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Petrophysical properties of the Kylylahti Cu-Au-Zn sulphide mineralization and its host rocks

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Abstract

This work presents a new set of petrophysical laboratory measurements from Kylylahti, a Cu-Au-Zn mine in the Outokumpu mining district, in the eastern Finland. Results are discussed and compared to earlier petrophysical data from the area. The study was aimed to provide a solid base for accurate interpretation of already existing geophysical exploration data and to new seismic data collected during the COGITO-MIN (COst-effective Geophysical Imaging Techniques for supporting Ongoing MINeral exploration in Europe) project.

The sample set covered the most common rock types found in Kylylahti. A small set of samples represented sulphide mineralizations from several mining sites in the Outokumpu district. In the area, ophiolitic ultra-mafic massifs consisting of Outokumpu assemblage rocks, are embedded in Kalevian sediments, black schists and mica schists. Several massifs, Kylylahti being one of them, contain polymetallic (Cu-Co-Zn-Ni-Ag-Au-Cd-Sn-As-Se-Mo) massive, semi-massive or disseminated sulphide mineralizations.

The petrophysical parameters measured were density, seismic P-wave velocity, porosity, magnetic susceptibility, intensity of remanent magnetization, inductive resistivity, galvanic resistivity and chargeability. Additional parameters calculated from the measurements were seismic impedance, Königsberger (Q) ratio and induced polarization (IP) estimates.

Density data divides the Kylylahti rocks in three categories: 1) Massive and semi-massive sulphide mineralizations with an average density of 3750 kg/m^3, 2) Outokumpu assemblage rocks with densities close to 3000 kg/m^3 and 3) Kalevian rocks with densities a bit under 2800 kg/m^3. Sulphide disseminations are common in Outokumpu assemblage carbonate-skarn-quartz rocks and black schists elevating the densities when abundant.

The average P-wave velocities for almost all Outokumpu assemblage rock types are a bit over 6 km/s. Soap stones, mica schists and black schists have lower P-wave velocities, around 5.5 km/s. Porosity of the samples was very low overall.

Most of the Kylylahti rocks belong to paramagnetic group (susceptibilities under 2000 µSI). Serpentinites and tremolitic calc-silicate rocks (TRECS) belong to strongly magnetic group as well as samples rich in disseminated sulphides. Low Q ratios reveal that magnetic mineral in serpentinites and TRECS is coarse-grained magnetite. Samples with disseminated sulphides have high Q ratios, thus the disseminations are mainly monoclinic pyrrhotites.

Both Kylylahti sulphide mineralizations and black schists are conductive as well as rocks rich in dissiminated sulphides. The rocks containing disseminated sulphides have high IP estimates. Conductivity of black schists is due to graphite and to some extent due to disseminated sulphides.

Physical properties of the ore samples from different mining sites reveal the differences in their mineralogy, mainly their changing proportions of pyrite, pyrrhotite and magnetite. The differences are due to metamorphic zoning in the Outokumpu district; the degree of metamorphism becomes higher when going from east to west or from surface to depth.

Recommended parameters, densities and P-wave velocities for seismic modelling in Kylylahti are given. Based on the results, the sulphide mineralizations should produce a detectable reflection against any background due to their high density. Also the other Outokumpu assemblage rocks have a clear contrast against the mica schists and black schists. Soap stones are an exception. The contact between Kalevian rocks and soap stones is hardly reflective at all, whereas soap stones in contact with other Outokumpu assemblage rocks form a reflecting contact.

Keywords

Petrophysics, Kylylahti, Outokumpu, COGITO-MIN, seismic modelling parameters

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Tiivistelmä

Tässä työssä esitellään uusia petrofysikaalisten laboratoriomittausten tuloksia Kylylahden Cu-Au-Zn kaivoksesta. Kylylahti sijaitsee Outokummun alueella Itä-Suomessa. Tuloksista keskustellaan ja niitä verrataan olemassaolevaan petrofysikaaliseen tietoon alueelta. Työn tarkoituksena oli tarjota vahva petrofysikaalinen pohja jo olemassaolevan geofysikaalisen aineiston sekä COGITO-MIN (COst-effective Geophysical Imaging Techniques for supporting Ongoing MINeral exploration in Europe) projektissa kerätyn uuden seismisen aineiston tulkintaan.

Työssä mitatut näytteet edustavat yleisimpiä Kylylahdessa tavattavia kivilajeja. Pieni osa näytteistä on malminäytteitä useammasta Outokummun alueen kaivoksesta. Alueella on Outokumpu-assosiaation kivistä koostuneita ofioliittisia ja ultramafisia massiiveja Kalevalaisten sedimenttien, mustaliuskeiden ja kiilleliuskeiden, ympäröiminä. Useat massiivit, Kylylahti mukaan lukien, sisältävät massiivisia, semi-massiivisia tai pirottuneita, monimetallisia (Cu-Co-Zn-Ni-Ag-Au-Cd-Sn-As-Se-Mo) sulfidimineralisaatioita

Näytteistä mitattiin tiheys, seismisen P-aallon nopeus, huokoisuus, magneettinen suskeptibilitetti, remanentin magnetoituman suuruus, induktiivinen ominaisvastus sekä galvaaninen ominaisvastus ja varautuvuus. Mitatuista arvoista laskettiin seisminen impedanssi, Königsbergin (Q) suhde sekä indusoidun polarisaation (IP) arvot.

Tiheyksien perusteella Kylylahden kivet voidaan jakaa kolmeen ryhmään: 1) massiiviset ja semi-massiiviset sulfidimineralisaatiot (tiheys keskimäärin 3750 kg/m^3), 2) Outokumpu-assosiaation kivet (tiheys noin 3000 kg/m^3) and 3) Kalevalaiset kivet (tiheys hiukan alle 2800 kg/m^3). Outokumpu-assosiaation karbonaatti-karsi-kvartsikivet sekä mustaliuskeet sisältävät yleisesti pirottuneita sulfidimineralisaatioita, jotka ollessaan runsaita nostavat kiven tiheyttä merkittävästi.

Keskimääräinen P-aallon nopeus Outokumpu-assosiaation kivissä on hiukan yli 6 km/s. Vuolukivissä, kiilleliuskeissa ja mustaliuskeissa nopeudet ovat matalampia, noin 5.5 km/s. Näytteiden huokoisuudet olivat hyvin matalia.

Useimmat Kylylahden kivet kuuluvat paramagneettiseen ryhmään (suskeptibiliteetti alle 2000 µSI). Serpentiniitit ja tremoliittiset kalkkisilikaatit (TRECS) kuuluvat vahvasti magneettiseen ryhmään, kuten myös näytteet, joissa on runsaasti pirottuneita sulfidimineralisaatioita. Serpentiniittien ja TRECSien matalat Q-suhteet kertovat, niiden sisältävän karkearakeista magnetiittia. Pirottuneita sulfidimineralisaatioita sisältävien näytteiden Q-suhteet ovat korkeita, joten pirotteet ovat pääosin monokliinistä pyrrhotiittiä.

Sekä Kylylahden sulfidimineralisaatiot että mustaliuskeet ovat sähköisesti johtavia, kuten myös pirotteisia sulfidimineralisaatioita sisältävät kivet. Viimeksimainituilla on korkeat IP-arvot. Mustaliuskeiden sähkönjohtavuuden aiheuttaa grafiitti sekä niissä olevat sulfidimineraalipirotteet.

Eri kaivosalueilta olevien malminäytteiden petrofysikaaliset ominaisuudet heijastavat eroja niiden mineralogiassa, pääosin pyriitin, pyrrhotiitin ja magnetiitin vaihtelevia osuuksia. Erot johtuvat Outokummun alueen metamorfisen asteen vaihtelusta. Metamorfinen aste nousee idästä länteen ja pinnalta syvemmälle mentäessä.

Työssä annetaan suositellut arvot tiheyksille ja P-aallon nopeuksille käytettäväksi Kylylahden seismisessä mallinnuksessa. Tulosten perusteella sulfidimineralisaatiot tuottavat havaittavan heijastuksen mitä tahansa taustaa vasten. Syynä on niiden korkea tiheys. Myös muiden Outokumpu-assosiaation kivien seisminen kontrasti kiille- ja mustaliuskeisiin on selvä. Vuolukivet muodostavat poikkeuksen. Kalevalaisten liuskeiden ja vuolukivien kontaktit eivät ole juuri lainkaan heijastavia, kun taas vuolukivien ja muiden Outokumpu-assosiaation kivien välinen kontakti tuottaa selvän heijastuksen.

Avainsanat
Petrofysiikka, Kylylahti, Outokumpu, COGITO-MIN, seismisen mallinnuksen parametrit

Säilytyspaikka
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Contents

1 Introduction

Today's mineral exploration targets deeper seated ore bodies, because the most of the shallow ones have already been found. Direct access to deep rocks by drilling is expensive, thus geophysical exploration methods are becoming more and more important. Accurate interpretation of geophysical data relies upon accurate petrophysical characterization of the targets in order to find out if the targets can be detected by geophysical methods, and what kind of signature can be expected. For this, measurements are made in laboratories, or in-situ on outcrops and in boreholes. Both types of measurements have their stregths and weaknesses. They also complement each other.

This thesis was made as a part of the COGITO-MIN-project (COst-effective Geophysical Imaging Techniques for supporting Ongoing MINeral exploration in Europe), a three year project that started in January 2016. COGITO-MIN aims to develop cost-effective and novel geophysical deep mineral exploration techniques, with particular emphasis on seismic imaging. The research institutions and industry partners from Finland and Poland, University of Helsinki, Geological Survey of Finland (GTK), Institute of Geophysics in Polish Academy of Sciences, Boliden, Vibrometric and Geopartner, are collaborating on the project. The study area of the COGITO-MIN-project is Kylylahti Cu-Au-Zn mine in North Karelia, in the eastern Finland.

The Kylylahti mine is situated on the northeastern side of the Outokumpu mining and exploration district (Figure 1), one of the most important mining areas in Finland. Outokumpu district hosts several Cu-Co-Zn-Ni-Ag-Au semimassive to massive sulphide deposits, that occur in association with the alterated margins of serpentinized peridotite bodies. These rock assemblages hosting the Outokumpu-type sulphide deposits are called Outokumpu assemblage. The Outokumpu district has a long mining and exploration history since its discovery in 1910. Currently Kylylahti mine is the only operational sulphide mine in the area. It is operated by Boliden.

Kylylahti was chosen for the area studied in the COGITO-MIN-project for several reasons: the area has potential for deep seated ore bodies, there are previous geophysical and geological studies that provide a good starting point for a new project, and Kylylahti mine was willing to participate providing its own geophysical and geological data and expertise for the project. During the field phase of the project in the summer 2016, active and passive 2D and 3D surface seismic data, and borehole seismic data were collected. To characterize the petrophysical properties of the Kylylahti rock types, a set of rock samples was selected for laboratory measurements. Results of these measurements are presented in this work.

Petrophysical properties of the rocks in the Outokumpu district have been discussed most recently by Leväniemi (2016) and Tuomi (2016). Leväniemi has compiled and analyzed the density and the magnetic data available from the Outokumpu area. Tuomi discusses the density and P-wave velocity character-

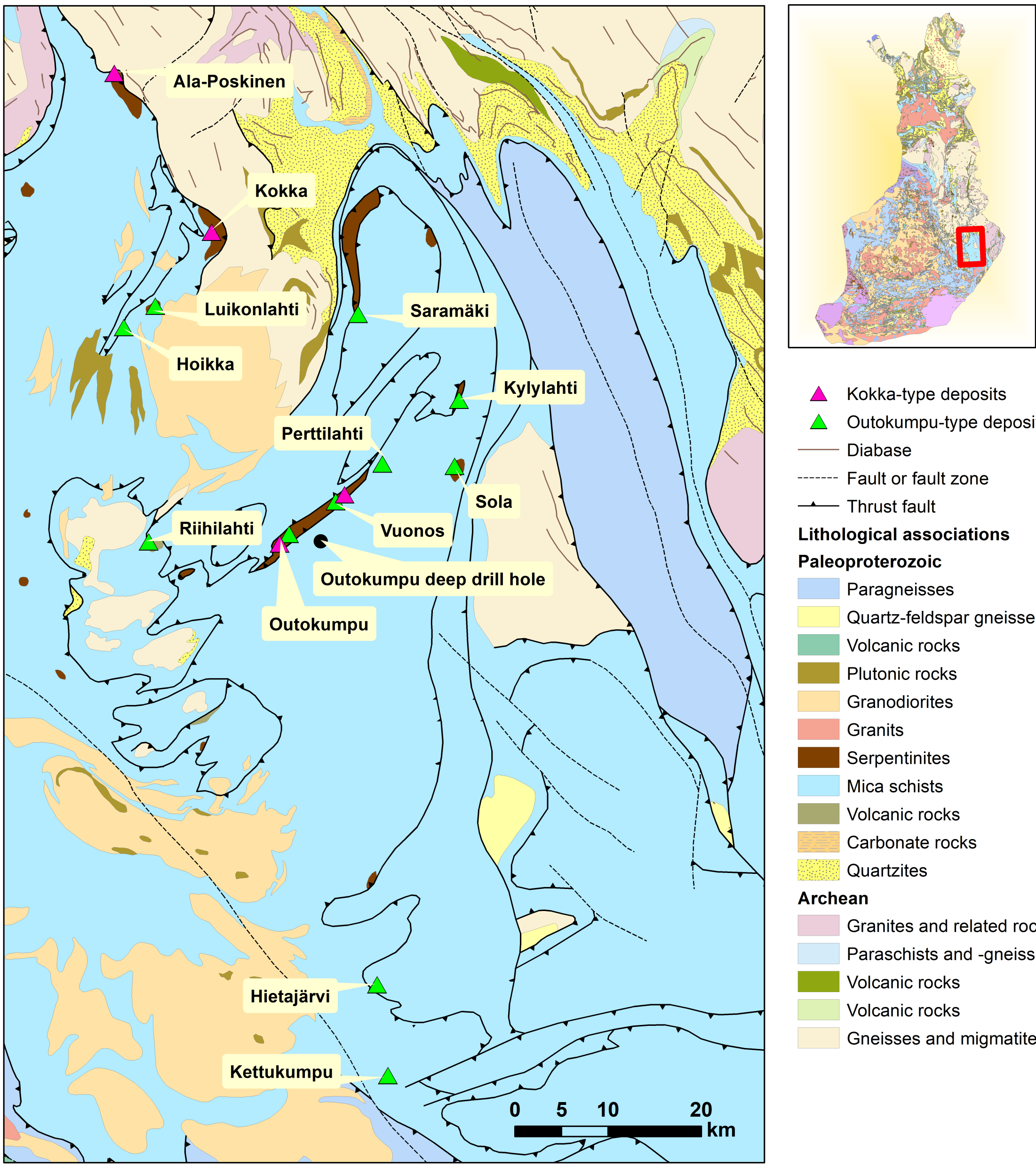

Figure 1: Generalized geological map of the Outokumpu district showing also the location of Outokumpu-type and Lokka-type sulphide deposits. Map modified from Generalized Bedrock of Finland ©GTK 2018 and Mineral Deposits ©GTK 2018.

istics. Several studies, e.g. Airo et al. (2011), Elbra et al. (2011) and Heinonen et al. (2011) present petrophysical data from the Outokumpu Deep Drill Hole core, including density, porosity, magnetic susceptibility, intensity of remanent magnetization, electrical properties (resistivity and chargeability), P-wave velocity, and thermal conductivity. Outokumpu Deep Drill Hole is a 2516 metres deep research bore hole drilled in 2004-2005. It is located about twenty kilometres southwest from Kylylahti (Figure 1). The Outokumpu Deep Drill Hole and core have been intensively studied from geophysical, geological and even biological points of view (Kukkonen 2011). Seismic wave velocity and velocity anisotropy of the Outokumpu Deep Drill Hole samples have been studied by Kern and Mengel (2011). Theoretical seismic velocity and density estimations for Outokumpu assemblage rocks and Outokumpu-type sulphide mineralizations have been made by Kukkonen et al. (2012b) and Tuomi (2016). Earlier, the petrophysical properties of the rocks in the Outokumpu dictrict have been discussed e.g. by Ketola (1973) and Ahokas (1984), Pekkarinen and Rekola (1994) and Ruotoistenmäki and Tervo (2006).

The emphasis of the COGITO-MIN-project is in the seismic imaging methods. During the planning phase it was concluded, that better constraints are needed to achieve more reliable seismic velocity values for the rocks in the Kylylahti massif. Earlier results (e.g. Heinonen et al. 2011, Kukkonen et al. 2012b) and recent seismic forward modelling results by Komminaho et al. (2016), imply that both the Outokumpu assemblage rocks and the sulphide mineralizations produce detectable seismic signals. However, seismic velocity measurements for the Outokumpu assemblage rocks have previously been available only from the Outokumpu Deep Drill Hole and Core (e.g. Airo et al. 2011, Elbra et al. 2011). In particular, no seismic velocity measurements were available for the Outokumpu-type sulphide mineralizations. Even worldwide, the seismic velocity data for chrystalline rocks is quite rare (e.g. Malehmir et al. 2013, and the references therein). The established variation of other petrophysical properties across the Outokumpu district (e.g. Leväniemi 2016), especially the density variation, implies that the reflectivity characteristics of the Outokumpu assemblage rocks and the Outokumpu-type sulphide mineralizations may vary. Thus, new seismic velocity and density measurements, as well as measurements of the other petrophysical properties, were needed for more accurate characterization of the targets.

The seismic reflection method has been used successfully already for a long time in soft rock environments, by oil and coal industries. In hard rock environments and within mining and exploration industry, other geophysical methods have traditionally been more attractive. Still, the only surface method that can provide high-definition images of the underground at depths required in metal exploration nowadays, is the seismic reflection method (Malehmir et al. 2012).

Sulphide mineralizations usually have high enough reflection coefficients against almost any background to be detectable. However, only big massive sulphide or iron deposits can be directly delineated by seismic methods. For example, a deposit in one kilometre depth should be more than 350 metres wide, and more than 20 metres thick, for it's dimensions to be accurately determined, assuming typical hard rock environment and dominant frequency of 100 Hz for the source. This is a typical resolution for a two-dimensional survey, with a vibroseis truck as a source. Better resolution can be achieved with higher frequencies (Salisbury and Snyder 2007, Malehmir et al. 2014). For example VIBSIST-200 system, used in COGITO-MIN VSP (Vertical Seismic Profiling) measurements, has a dominant frequency around 200 Hz (Riedel et al. 2017b).

Seismic methods have been successfully used in mine planning, and in mine site exploration (e.g. Malehmir et al. 2012, and the references therein). 2D and 3D surface or borehole reflection seismic data, together with other geophysical and petrophysical data from the area, can be used to produce reliable models of the lithological contacts and geological structures in the study site. This information can be used to guide the choises of other exploration methods, especially drilling. Structural information can also be used in geotechnical planning of the mine (e.g. Koivisto et al. 2015).

The aim of this work was to deepen the knowledge of petrophysical parameters in the COGITO-MIN study area, and to provide a solid base for accurate interpretation of already existing geophysical exploration data and to new seismic data collected during the COGITO-MIN-project. Further, this study was intended to provide parameters needed in the seismic forward modelling of COGITO-MIN seismic data. Petrophysical differences between the Outokumpu-type sulphide mineralizations from the entire Outokumpu district are also investigated through a small set of ore samples from several Outokumpu district mining sites.

First, the geology and geophysical setting of the Outokumpu mining and exploration district, and the Kylylahti mine area are presented. Next, the datasets used in the analyses are explained. After that, the laboratory measurement methods and measurement practices are described. Then the results of the petrophysical laboratory measurements are presented and the results discussed and compared to earlier petrophysical data from the area. Finally, implications of the results for the interpretation of geophysical data from the area, in particular seismic reflection data are discussed. As a conclusion, recommended petrophysical parameters for seismic modelling in Kylylahti will be given, and suitability of different geophysical methods for exploration in Kylylahti will be estimated.

2 Geological and geophysical setting

2.1 General geological setting

The geological setting of the Outokumpu district can be seen in Figure 1. The Outokumpu district forms a part of the North Karelian schist belt, located between the Archean Karelian Craton in the northeast, and the Paleoproterozoic Svecofennian domain of accreted island-arc terrains in southwest. The North Karelian schist belt consist mainly of Jatulian (2.5-2.0 Ga old) and Kalevian (2.0-1.9 Ga old) metasedimentary rocks, with some intercalations of tholeiitic volcanites. After the deposition of pre-orogenic Jatulian sediments, the rifted western margin of the Karelian Craton became eroded, forming a clear unconformity between the Jatulian and Karelian sequences. Kalevian sediments mark the beginning of the Svecofennian orogenic development. A passive continental margin was formed, and Kalevian sediments were deposited on the new, intensely faulted continental margin (Gaál and Gorbatschev 1987).

The Outokumpu district lies within an allochthonous nappe complex, that was thrusted onto the Karelian Craton margin during the early stages of the Svecofennian orogeny. Kalevian mica schists resting upon Archean gneiss granitoid basement are dominant in the area. Ophiolitic ultramafic rocks are enclosed within the mica schists. A sequence of these ophiolitic bodies, containing Outokumpu, Vuonos and Kylylahti mines, is called the Outokumpu Belt. The Archean basement is revealed on surface north and northeast of the Outokumpu district. Between the Archean basement and the overlying Kalevian strata, in the northern parts of the Outokumpu district, is a sheet of Jatulian quartzites and arkoses. The quartzite sheet is intruded by ultramafic-mafic sills up to hundreds of metres thick, that are also common in the Archean basement, close to the Jatuli-Archean interface. To the west of the Outokumpu Belt, the Archean basement is intruded by the 1.86 Ga Maarianvaara granitoid suite (Kontinen et al. 2006, Peltonen et al. 2008). Since the cooling of the eastern Finland after the Svecofennian orogeny no significant tectonical events have occurred in the area, only erosion has levelled the ground to its current surface level and several glaciations have formed the landscape (Lehtinen et al. 1998).

The geological structure of the Outokumpu district is complicated. The vast majority of its rocks are Kalevian metasediments. The ophiolite complex consists of hundreds of individual massifs and fragments distributed over an area of more than 5000 km^2, as can be seen in Figure 1. In general, the rocks in the Outokumpu district are highly strained and metamorphosed (Kontinen et al. 2006).

The Kalevian strata in the Outokumpu district is divided into parautochthonous Upper and autochthonous Lower Kaleva. Both comprise mainly metaturbiditic greywackes, mica schists. In the Lower Kaleva there are thin intercalations of tholeitic metabasalts and black schists, whereas in the Upper Kaleva there are thick intercalations of black schists. In the basal parts of the unit they enclose fragmented ophiolite bodies, mainly serpentinites, and the associated Outokumpu-type, semi-massive to massive sulphide deposits (Kontinen et al. 2006, Peltonen et al. 2008). In a typical set-up an ultramafic massif consisting of Outokumpu assemblage rocks is enveloped by black schists, and hosted by mica schists. All the known Outokumpu-type sulphide deposits (Figure 1), except for Riihilahti, are found in association with ultramafic bodies. Outokumpu assemblage rocks are also found as smaller layers and lenses in the mica schist outside of the massifs. The ultramafic massifs contain small amounts of deformed and metamorphosed basaltic rocks, mainly as small stocks, sills or dykes (Koistinen 1981, Kontinen 2005, Peltonen et al. 2008).

2.2 Exploration and mining history of the Outokumpu district

The first indication of a sulphide mineralization in the Outokumpu district was the Kivisalmi boulder, an erratic glacial boulder found in 1908 in Rääkkylä, about 50 kilometres southeast from Outokumpu (Saksela 1948). The source of the boulder, the uppermost part of the Outokumpu ore, was found in 1910. The Outokumpu mine was operational between the years 1913 and 1989 (Kontinen et al. 2006). At the time it was one of the biggest copper producers of the world. It can be said, that the discovery of the Outokumpu ore was the starting point for development of the modern metal industry in Finland.

Three biggest Outokumpu-type deposits, Outokumpu, Vuonos and Luikonlahti, have been mined away. The total production of these mines was approximately 50Mt of ore, averaging 2.8 wt.% copper, 1 wt.% zinc, and 0.2 wt.% cobalt, together with minor amounts of nickel and gold. In addition to these sites, several sub-economic Outokumpu-type mineralizations and Ni mineralizations, called Kokka type, have been found in the Outokumpu district. The most recently found economically significant Outokumpu-type deposit is Kylylahti (Kontinen 2005, Peltonen et al. 2008). Mining in Kylylahti was started in 2011. The known deposits in the Outokumpu district are shown in Figure 1, and grades and tonnes of Outokumpu-type ores investigated in this work are shown in Table 1.

Table 1: Grades and tonnes of selected Outokumpu-type ore deposits

Deposit	Cu wt.%	Co wt.%	Zn wt.%	Ni wt.%	Au wt.%	Ag wt.%	Fe wt.%	S wt.%	tonnes Mt
Outokumpu	3.8	0.24	1.07	0.12	0.8	8.90	28.11	25.3	28.5
Vuonos	2.45	0.15	1.6	0.13	0.1	11	24.8	17.5	5.89
Kylylahti	2.63	0.39	0.76	0.13	0.9	-	-	20.6	1.95
Luikonlahti	0.99	0.11	0.5	0.1	-	-	25	16.5	7.5

Data from (Peltonen et al. 2008).

The Vuonos deposit was found in 1965 by following the Outokumpu-type rock assemblage from Outokumpu to the northeast, and utilizing trends in its Co/Ni ratio for targeting the exploration drilling. The deposit was mined between 1972 and 1986 (Kontinen et al. 2006).

The Luikonlahti deposit is not a clear continuation of the same geological formation than Outokumpu and Vuonos, as can be seen from Figure 1. It is situated on a different branch of the ophiolite complex. First signs of a sulphide mineralization in Luikonlahti were found already in the 1910's, after the discovery of the Outokumpu ore, but the deposit turned out to be economically important only as late as 1958. The decision to open a mine and a concentration plant at Luikonlahti was made in 1965. The first batch of copper concentrate was produced in 1968. Mining at Luikonlahti came to an end in 1983, but its processing plant serves the Kylylahti mine even today (Kontinen et al. 2006).

The Kylylahti deposit was found in 1984 as a result of intensified exploration efforts in the Outokumpu district before the shut down of the Outokumpu and Vuonos mines (Kontinen et al. 2006). Mining in Kylylahti was started in 2011 by Altona Mining Ltd, and the mine has been operated by Boliden since 2014. At the end of the year 2016 the total mineral reserve has been 1.9 Mtonnes, and the reserve life at full production is estimated to last about 2 years (Boliden 2016a, "Kylylahti Mineral Deposit Report" 2016).

2.3 Outokumpu-type sulphide mineralizations and their physical properties

The Outokumpu-type sulphide mineralization is a rare type of Precambrian, hydrothermal sulphide mineralization related to ultramafic rocks. No deposit outside of the Outokumpu district has been unambiguously categorized as of Outokumpu-type (Rasilainen et al. 2014). The three most similar sulphide deposits are Dur'ngoi massive Cu-Co-Zn sulphide deposit in China, Eastern Metals Ni-Cu-Zn-deposit in Canada, and Sykesville Fe-Cu-Co-Zn-Ni deposit in the USA. The best modern analogues for Outokumpu-type sulphide mineralization can be found from peridotite-floored massive sulphides in mid-ocean ridges, although these seafloor sulphides contain much less nickel than the Outokumpu-type mineralizations (Peltonen et al. 2008).

According to Peltonen et al. (2008), an Outokumpu-type mineralization has to meet the following criteria. The deposit is located within an accreted terrain. The sulphide mineralization is intimately associated with residual upper mantle peridotites. No hydrothermal-exhalative sedimentary rocks should be present. The mineralization consists of polymetallic (Cu-Co-Zn-Ni-Ag-Au-Cd-Sn-As-Se-Mo) massive or disseminated sulphides. The mineralization has extremely low lead abundance, a mantle-like lead isotopic signature, and abnormally high nickel content. Sulphur and metals should be derived from both mantle- and crustal-like sources, and there should have been syntectonic remobilization and upgrading of the sulphides.

The deposit modelling of the Outokumpu-type sulphide mineralizations has been under active research since 1920s. Peltonen et al. (2008) summarizes the results so far, and proposes a model that best explains the features seen today in the mineralizations of the Outokumpu district. The model has three stages (Figure 2).

At first, in the ocean floor stage, copper proto ore was formed in an ultramafic ocean floor environment, as a result of hydrothermal convection, caused by extensional faulting and hot magmas. This occured about 1950 Ma ago (Peltonen et al. 2008).

The second stage, the obduction stage, started in the beginning of the Svecofennian orogeny, circa 1910 Ma ago, when the Svecofennian island arc was docked to the Karelian craton. The Outokumpu nappes were thrusted onto the craton, folded, and elongated. The regionally prominent schistosity of rocks was developed. The originally oval-shaped copper proto ore plates became narrow, long ribbons, parallel with the fold hinges. At the same time nickel proto ore was formed in the obducted sea floor, in a low temperature (100-200°C) listwaenite-birbirite-type carbonate-quartz alteration of peridotite body margins (Koistinen 1981, Kontinen et al. 2006, Peltonen et al. 2008).

In the third stage, the regional deformation stage, circa 1880 Ma ago, proto ores blended, when copper end-member sulphides were hydrothermally mobilized, and emplaced to nickel sulphide bearing quartz-carbonate rocks. The final emplacement of mixed ore bodies was structurally controlled, leading to somewhat different setups in different parts of the Outokumpu district. However, the ore bodies are everywhere located along or near the interface between black schists and alterated peridotite body margins. The final emplacement took place at the late stages of the regional tectonic activity, but before the peak of the regional metamorphism (Peltonen et al. 2008). The metamorphic peak was attained in a relatively late stage of the Svecofennian orogeny, about 1860 Ma ago. In the east, the thermal peak took place in low amphibolite facies (500-550 °C), and in the west in upper amphibolite facies (700-770 °C), at pressures of 2-4 kb. The pressure peak was reached already in the early stage of the orogeny (Kontinen et al. 2006).

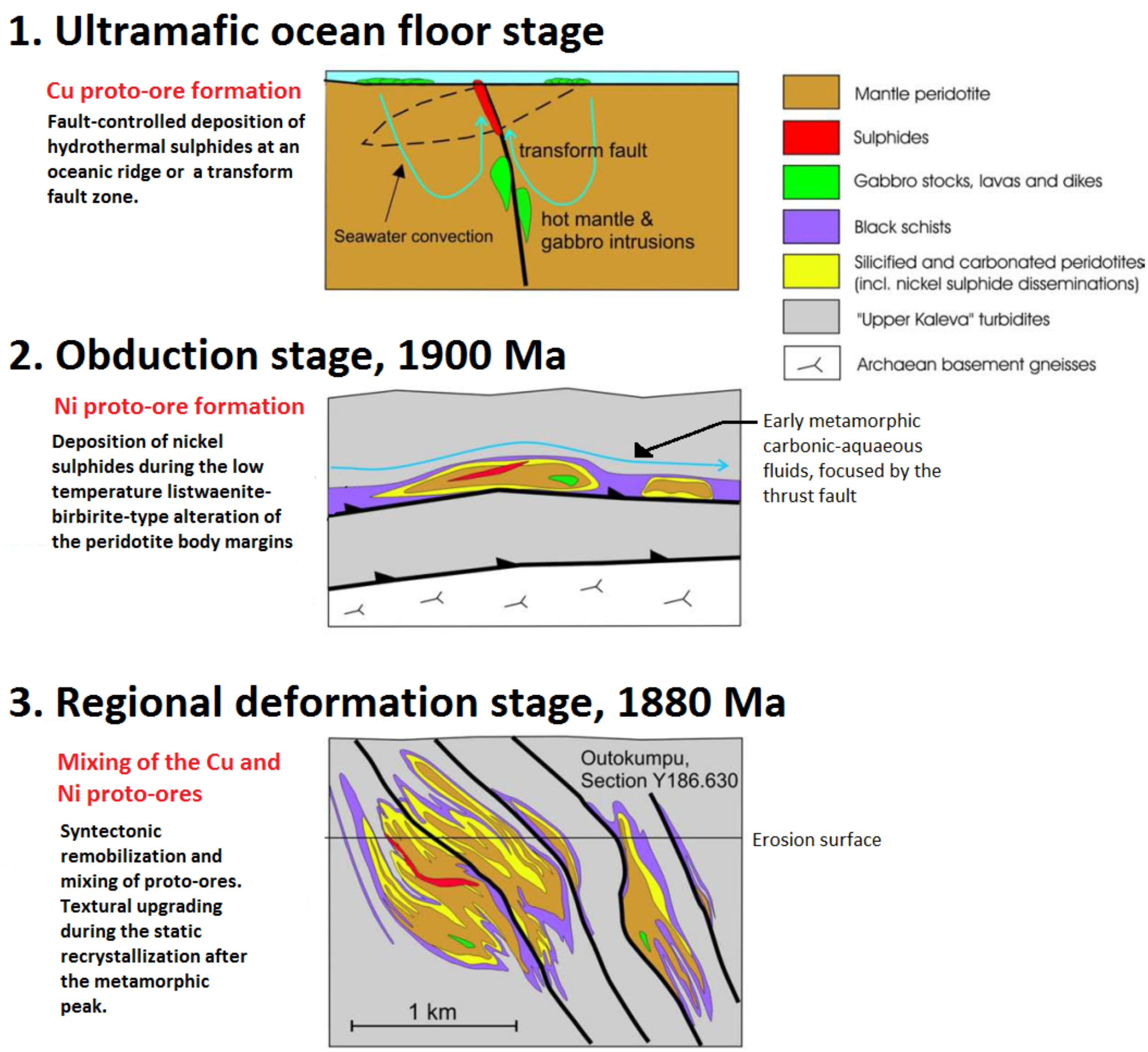

Figure 2: A cartoon of the mineral deposit model for Outokumpu-type sulphide deposits and Ni disseminations illustrating the three main stages of the sulphide mineralization genesis. Modified from (Peltonen et al. 2008).

Mineral deposits are often classified according to their genetics. For the purposes of geophysical exploration, this is the most useful classification scheme as the ore forming processes are largely reflected in differencies of the physical properties of the mineralization and the host rock. A classification based only on metal content would not be sufficient, because most metals show quite similar physical properties, and are thus not distinguishable (Airo 2015). When choosing or developing geophysical exploration methods, it is always beneficial to have references. Outokumpu-type deposits are very much like volcanogenic massive sulphide (VMS) deposits, in particular when considering their physical properties, and are often discussed jointly (e.g. Weihed et al. 2005, Galley et al. 2007, Airo 2015). The main difference between these two deposit types is that VMS deposits form in hydrothermal-exhalative processes at or near the sea floor (Galley et al. 2007), but the Outokumpu-type deposits lack hydrothermal-exhalative sedimentary rocks, and are instead related to ophiolitic rocks (Kontinen 1998, Peltonen et al. 2008). However, because the Outokumpu-type deposits are so rare, the use of VMS type deposits as a reference is justified.

The physical properties of VMS and Outokumpu-type deposits are in a marked contrast to their host rocks, and are therefore well suited for geophysical exploration (e.g. Moon et al. 2006). Airo (2015) summarizes the physical properties of sulphide mineralizations. Because of their high density values, gravity surveys are often successful. Sulphide deposits are usually conductive compared to their host rocks, and thus electromagnetic methods are effective. However, if the sulphide minerals are disseminated, electromagnetic methods may not be suitable. In such cases, induced polarization can be used to detect disseminated sulphides. Magnetic methods might not reveal the deposit itself, but are useful in identifying broad geological framework: lithological contrasts, crustal structures, and type and degree of alteration. Sulphide deposits are also often detectable in reflection seismics, due to their high seismic impedances (Airo 2015).

2.4 Rock types

The rock naming in this work, and practically in all recent studies concerning the Outokumpu district, is done according to the unified naming practise developed during the GEOMEX project, years 1998 to 2003. GEOMEX was a joint project of the Geological Survey of Finland and Outokumpu Mining Oy. It was launched to conduct geological and geophysical exploration and modelling in the Outokumpu district, and to collect, archive, and revise existing data and drill cores (Kontinen et al. 2006, Ruotoistenmäki and Tervo 2006). Before GEOMEX, the naming protocols were non-coherent, making any computer-based application difficult (Kontinen 2005). The lithological groups, lithologies and the rock types within them, that are used in this work, are compiled in Table 2. GEOMEX classification as a whole contains eight main lithologic groups, with tens of lithologies and rock types (GTK 2016, written comm.).

In this work we have rock types from two lithological groups: Outokumpu assemblage and Schists after sedimentary rocks. Outokumpu assemblage is divided into three lithologies: Outokumpu ultramafic and derived rocks (OUM), Outokumpu metabasites (OBA) and Sulphide mineralizations (SULMI). Each of these lithologies contain several rock types. In Table 2 are listed only rock types, that are represented in the sample set of this work. From the Schists after sedimentary rocks we have samples from lithologies Wackes-muds (WA), Black Schists (BS), Carbonate rocks and marls (CRB). Also Iron formation (IF) is listed, because it will appear in the drill core data of the chapter 5.6.

In this work, rocks rich in sulphide mineralizations have been called disseminated, massive or semi-massive ore. As ore is an economical, rather than a geological or geophysical term, it is recommended to use the term sulphide mineralization, except when discussing an ore of a mine. All the sulphide mineralization samples studied in this work originate from an ore in a mine and thus their original naming has been conserved. Otherwise, the term sulphide mineralization is used, if not the ore of a certain mine is discussed.

Table 2: **Lithological groups, lithologies and rock types used in this work**
Abbreviations are given after the names

Outokumpu assemblage
Outokumpu ultramafic and derived rocks OUM
- Serpentinite SP
- Soap stone SS
- Carbonate rock CRBR
- Skarn SKA
- Tremolitic skarn TRESKA
- Quartz rock QTZR
Outokumpu metabasites OBA
- Chlorite Schist CHLS
Sulphide mineralizations SULMI
- Massive ore MS
- Semi-massive ore SMS
- Disseminated ore DIS
Schists after sedimentary rocks
Wackes-muds WA
- Mica schist MCAS
Black Schists BS
- Black schist BS
- Sulphide-rich black schist SULBS
Carbonate rocks and marls CRB
- Carbonate rock CRB
- Tremolitic calc-silicate rock TRECS
Iron formations (IF)
- Oxide iron formation OXIF
Source: GEOMEX rock classification and names (GTK 2016, written comm.)

2.4.1 Serpentinites and soapstones

Serpentinites (SP) and soapstones (SS) form the biggest volume of rock in the ultramafic, ophiolitic bodies of the Outokumpu district. The chemical composition of these rocks corresponds to mantle peridotites, lhertzolites, harzburgites, and dunites, implying a depleted mantle origin (Peltonen et al. 2008). The present mineral compositions and textures reflect the regional east-west metamorphic zoning. These rocks were already serpentinized before the latest regional metamorphism (Säntti et al. 2006).

During the latest regional metamorphism in the eastern parts of the Outokumpu district, in lower amphibolite facies, the serpentinites became metaserpentinites, mainly consisting of fine-grained antigorite, with some carbonate and talc. Locally, usually at the margins of the bodies, antigorite has completely been replaced, and soapstones were produced. In GEOMEX classification, rocks are classified as soap stones when they contain more than 95 % carbonate and talc. Soap stones are often also called talc-carbonate rocks. In the core parts of the massifs, metamorphic grade is higher, and antigorite is gradually replaced by olivine and tremolite porphyroblasts. When moving towards west, the metamorphic grade changes to middle and then to upper amphibolite facies. Antigorite is first gradually replaced by olivine and/or tremolite porphyroblasts, then by talc and olivine, after which by anthophyllite-olivine, and finally by enstatite-olivine. These western rocks are metaperidotites, that have commonly hydrated to lizardite-chrysotile serpentinites after the peak of metamorphism (Peltonen et al. 2008).

Idealized rock columns, showing the lithology of the altered margins of the ultramafic bodies of Outokumpu district, are shown in Figure 3 for the lower and the higher grade metamorphic zones. These columns show the differences in the serpentinite parts of the bodies. Metaserpentinites and metaperidotites both are called serpentinites, unless the differences within this rock group will be discussed. The same naming policy has been used in the GEOMEX project (Kontinen et al. 2006).

Typically the Outokumpu serpentinites lack preferred orientation. They contain coarse-grained chromite grains, as is usually the case with mantle peridotites. Composition and texture of chromites vary with metamorphic grade (Peltonen et al. 2008).

2.4.2 Carbonate, skarn, and quartz rocks

Carbonate (CRBR), skarn (SKA), and quartz rocks (QTZR) of the Outokumpu assemblage are interpreted as products of silification and carbonation of the outer margins of the peridotite massifs, during or immediately after the obduction of the seafloor (Peltonen et al. 2008). In the GEOMEX classification, rocks comprising more than 50 vol% carbonate, are defined as carbonate rocks. Those with more than 50 vol% tremolite are tremolite skarns, and rocks with more than 50 vol% quartz are quartz rocks. Gradational and complex mixing relationships of Outokumpu carbonate-skarn-quartz rocks make them difficult to classify consistently. When re-analyzed with X-ray fluorescence method (XRF), it appeared that tens of samples were misnamed (Kontinen 2005). Consequently, this group of rocks is often discussed as one rock type. As is the case with almost all the rocks in the Outokumpu district, also the composition of carbonate, skarn, and quartz rocks reflects the metamorphic zonation of the area (Säntti et al. 2006).

The carbonate rocks in the Outokumpu assemblage are generally found as thin, 1-5 m thick, discontinous seams between serpentinite and skarn-quartz rocks. The main mineral constituents are dolomite, calcite, and locally magnesite and tremolite. At increased amount of tremolite, the carbonate rocks

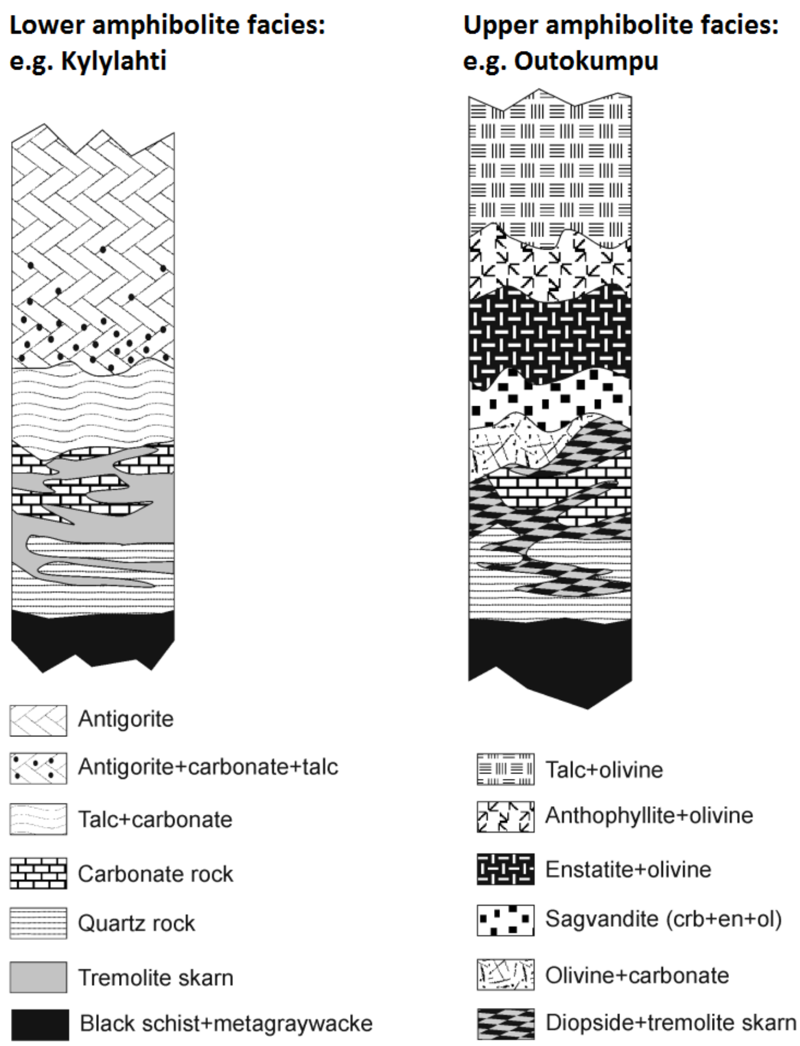

Figure 3: Idealized columns depicting the mineralogical and lithological successions of the metasomatically altered margins of the metaserpentinite bodies in lower amphibolite facies metamorphic zone (e.g. Kylylahti) and metaperidotite bodies in upper amphibolite facies metamorphic zone (e.g. Outokumpu). Modified after (Säntti et al. 2006)

grade into tremolite-carbonate rocks, tremolite skarns. Towards metaperidotites, the carbonate rocks may grade into olivine carbonate rocks and sagvandites (Figure 3). Chromite or a weak to modest dissemination of iron sulphides and pendlandite are quite common (Peltonen et al. 2008). It should be noticed that there are also Kalevian carbonate rocks (CRB) of sedimentary origin in the Outokumpu district.

Calc-silicate rocks, skarns, are typically situated between carbonate and quartz rocks, or as band layers inside quartz rocks. In the lower metamorphic zone, these are tremolite-carbonate skarns and in the higher metamorphic zone, diopside-tremolite skarns (Figure 3). The main minerals are tremolite, dolomite, and/or calcite, and quartz. In the upper amphibolite metamorphic environments skarns can also contain olivine. Common minor minerals include chromite, disseminated iron sulphides, pentlandite, and graphite. A local specialty is eskolaite, an uncommon chromium oxide, first found in Outokumpu in 1949, and named in the honor of Finnish geologist Pentti Eskola (Kouvo and Vuorelainen 1958, Peltonen et al. 2008).

Quartz rocks in the Outokumpu assemblage are fine-grained and strongly schistose. These are quartz-tremolite and/or diopside banded rocks, which contain some chromite. Pyrrhotite ± pyrite ± pentlandite disseminations as speckles or stripes are also common. Quartz rocks show gradation into tremolite skarn, and near to black schists they contain abundant fine-grained graphite (Peltonen et al. 2008).

2.4.3 Outokumpu metabasites

Chlorite schists (CHLS) of the sample set of this work belong to Outokumpu metabasites, which are mafic, basaltic rocks forming 5-25% of the Outokumpu district ultramafic bodies. Outokumpu metabasites occur mostly as dykes or small stocks. They have apophyses and chilled margins against the enclosing ultramafic rocks dating them back to the time, when the ultramafic rocks were still a part of the ocean floor (Peltonen et al. 2008).

Outokumpu metabasites are strongly schistose and folded metagabbros and amphibolites. Narrow dykes are usually pervasively altered to chlorite schist, but occasionally have metagabbroic or amphibolitic cores. Also the chloritized parts are schistose, showing that they were already chloritized before the peak of metamorphism. The main mineral constituents of metagabbros are hornblende, plagioclase, and epidote/clinozoisite, with minor traces of ilmenite and sulphides (Kontinen et al. 2006).

2.4.4 Sulphide mineralizations

Outokumpu-type Cu-Co-Zn-Ni-Ag-Au sulphide mineralizations occur typically as thin, narrow, and sharply bounded sheets, lenses, or rods between alterated fringes of ultramafic massifs and black schists surrounding the massifs. Outokumpu-type mineralizations frequently contain also minor tin, arsenic, and selenium, while metals such as bismuth, antimony, and lead are usually present only in very low quantities, even in the most metal-rich parts of the deposits (Kontinen et al. 2006).

In addition to the massive (MS) and semi-massive (SMS) sulphide mineralizations, Outokumpu assemblage skarn-quartz rocks host semimassive and disseminated, subeconomic nickel mineralizations (DIS), called Kokka-type Ni-mineralizations. Some of these are next to the massive-semimassive sulphide mineralizations, but many are far from those and apparently unrelated with any Outokumpu-type sulphide mineralizations. Kokka-type Ni-mineralizations are characterized by very low copper and cobalt concentrations and mantle-like Co/Ni ratios. They have clearly different origin than the Outokumpu-type mineralizations, but are probably the source of high nickel concentrations in some of the Outokumpu-type mineralizations (Kontinen et al. 2006, Peltonen et al. 2008).

Geochemical and mineralogical analyses of Outokumpu district sulphide mineralizations show differences between the different sites (e.g. Kontinen et al. 2006). The sulphide mineralizations from Kylylahti, Outokumpu, Vuonos, and Luikonlahti will be discussed in this work. The biggest difference between these sites is in the metamorphic grade. According to the growing grade of metamorphism from east to west and from surface to depth, the peak metamorphic temperatures were the highest in Luikonlahti and the lowest in Kylylahti (e.g. Säntti et al. 2006). This is clearly seen on the other rocks of Outokumpu assemblage, but can also be seen in the sulphide ratios of the sites. In Kylylahti, the sulphide mineralization is pyrite dominated, becoming pyrrhotite dominated in the deeper parts of the deposit. Also magnetite becomes more abundant with depth and increasing grade of metamorphism. On the western side, Luikonlahti is pyrrhotite dominated and rich in magnetite. Pyrite exists only as a minor sulphide. Vuonos is pyrrhotite dominated, while in Outokumpu large amounts of pyritic sulphide mineralization can be found (Kontinen et al. 2006).

2.4.5 Mica schists and black schists

A major part of the North Karelian Schist belt, and thus also of the Outokumpu district, is made of metasedimentary rocks. The most common of these is mica schist (MCAS). Mica schists are intercalated by black schists (BS). Black schist formations vary from a few centimetres to some metres thick. The ultramafic massifs in the Outokumpu district are usually associated with thick black schist layers. Also carbonate (CRB) and calc-silicate rocks (TRECS), with origins in sedimentary rocks, can be found. These differ from their peridotite-derived counterparts by having lower chromium contents (Kontinen 2005, Kontinen et al. 2006).

Outokumpu district mica schists are from medium to fine-grained, mineralogically simple biotite-quartz-plagioclase schists. These derive mainly from immature, muddy sands deposited usually in 10 to 120 cm thick tabular turbidite beds. Where sedimentary structures are preserved, sandy beds show sharp, erosional bases. Intervening mud-shale layers are very thin or lack totally. Outokumpu district mica schists are quartz-intermediate with SiO_2 between 68 and 74 wt.%, while the intercalated mud-shales have 53-65 wt.% of SiO_2 (Kontinen et al. 2006).

According to the Finnish field geological practise, any fine-grained, siliclastic metasedimentary rocks containing more than 2 wt.% graphitic carbon are regarded as black schists, whether these are black metashales or sandy metamuds (Kontinen 2005). Outokumpu district black schists have previously been classified as argillaceous, calcareous, and arenaceous types, depending on their clay, amphibole or quartz content, respectively (Kontinen et al. 2006). For the purposes of this study, it is sufficient to distinguish sulphide-rich black schists (SULBS) from ordinary black schists (BS).

Black schists have been popular rock types for exploration because of their spatial association with the Outokumpu-type mineralizations, and their strong enrichment in many metals, such as Ag, Au, As, Co, Cu, Mo, Ni, V, Se, Zn and U. Unfortunately detailed chemical and lithological studies have found no evidence of any tie between black schists and Outokumpu-type sulphide mineralizations. Outokumpu district black schists are sedimentary rocks intercalated with mica schists, mineralogically ressemble mica schists, and are therefore probably coevally deposited and have a common provenance (Kontinen et al. 2006), whereas the origin of the Outokumpu-type sulphide mineralizations lies already in the time when the ophiolitic bodies were still a part of the sea floor (Peltonen et al. 2008).

2.5 Characteristics of the Kylylahti deposit

The Kylylahti massif, containing the Kylylahti deposit, is located at the nose of the Outokumpu nappe (Figure 1). The massif is shown in detail on the geological map in Figure 4. Kylylahti massif has been interpreted as a tight and upright folded, 100-200 m thick package of tectonically stacked rocks. Its geological structure is complicated and can be interpreted in other ways too (Kontinen 2005). The Kylylahti deposit is at the eastern margin of the Kylylahti massif, in a prominent late fault or shear zone. It's location in the fault zone implies that pervasive syntectonic reworking of the sulphides has taken place, in agreement with the deposit model of Peltonen et al (2008). Kylylahti lies within the lowest grade metamorphic zone in the Outokumpu district, which is clearly seen when the rocks of Kylylahti are compared to rocks from other parts of the Outokumpu district (Säntti et al. 2006).

The terrain in Kylylahti is small-patterned, hummocky, moraine covered landscape, with small bogs and a few ponds. The quartenary moraine cover is 5-25 m thick. There are only a few outcrops in the area. The submoraine bedrock in Kylylahti can be seen in the geological map in Figure 4. In contrast to the often unweathered and ice-polished, hard bedrock commonly found in North Karelia, Kylylahti area has several peculiar pothole-like depressions of weathered bedrock, filled by fine-grained micaceous-talcose material, which seems to originate from in-situ weathering of the bedrock. Underneath these depressions, the rock type is usually soapstone (Kontinen 2005).

The Kylylahti massif is substantial by its area and volume, but it is poorly exposed. It consists of several imbricated, sheared bodies of serpentinite and soapstone, which in many places have thin marginal selvages of carbonate-skarn-quartz rocks. An example of the internal structure of the Kylylahti massif can be seen in Figure 5. In Kylylahti, serpentinites are mostly fine-grained and massive antigorite rocks. Carbonate porphyroblasts and talc flakes replacing antigorite are common. Chromite and its oxide alteration products are frequent minor constituents. Olivine porphyroblasts appear to serpentinites and soapstones circa 600 m below the surface, indicating increasing grade of metamorphism towards the depth. Kylylahti serpentinites have relatively high sulphur content but they are not significantly enriched in copper, cobalt, or nickel except for a weak oxide-sulphide mineralization inside of the serpentinites, at the hanging wall of the deep sulphide disseminations (Kontinen 2005).

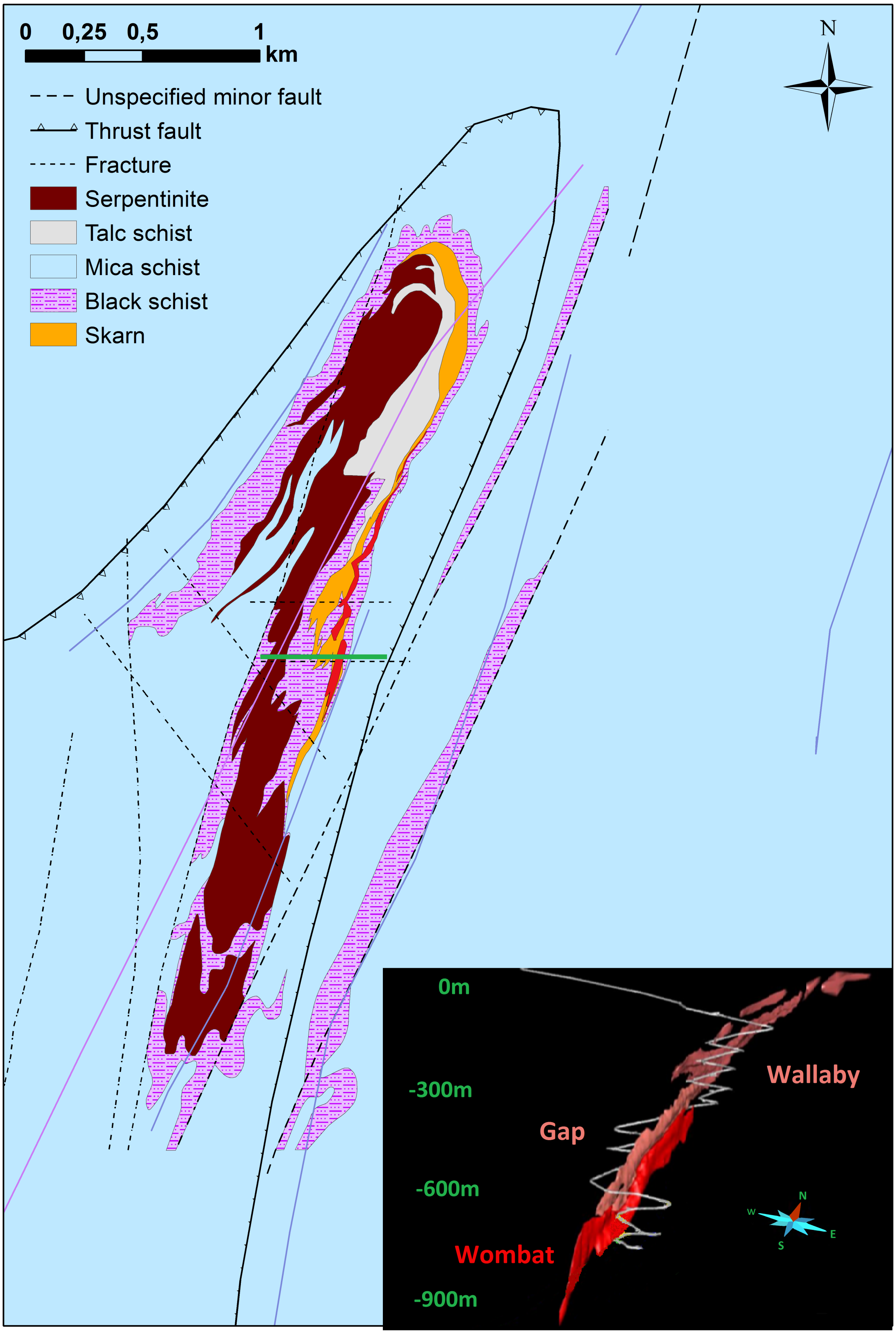

Figure 4: Geological map of the Kylylahti massif, showing surface projection of the ore (on red) and location of the cross-section profile (on green), which is presented in Figure 5. Insert shows a side view of the Kylylahti ore towards north. The mine tunnel is shown on white. Modified from Bedrock of Finland ©GTK 2018.

The Kylylahti massif margins are usually graded from antigorite serpentinite to soapstone within a few metres. Kylylahti soapstones consist of fine-grained talc hosting carbonate porphyroblasts. Talc to carbonate ratio is typically 1/1, but with variation up to pure talc schist. Common accessory minerals include pyrrhotite, pentlandite, gersdorftite, and chromites. As is the case with the serpentinites, also the soapstones in Kylylahti are quite rich in sulphur, but do not have any obvious copper, cobalt, or nickel enrichments (Kontinen 2005). Kylylahti soapstones have been mined during the 1970s, and still there is a considerable talc potential left. Six kilometres to the southwest of Kylylahti, on another serpentinite-soapstone massif, is Horsmanaho talc mine, operated by Mondo Minerals (Kontinen 2005, GTK 2016).

The alterated zones of the Kylylahti massif contain typical Outokumpu assemblage carbonate-skarn-quartz rocks. Quartz rocks do not contain uvarovite or chromian diopside, which are characteristic for quartz rocks in other parts of the Outokumpu district. This is because the metamorphic grade in Kylylahti was too low for formation of these minerals. Along the mineralized eastern side of the massif, a significant part of the skarn material occurs in veins and complex breccia-like networks of massive carbonate-tremolite skarn. These skarns cut and replace Outokumpu-type carbonate-skarn-quartz rocks and black schists, and are thus younger in origin. These younger skarns are the principal host of Cu-Co-Zn disseminated sulphide mineralizations in Kylylahti (Kontinen 2005).

The Kylylahti massif contains 5-10 vol.% Outokumpu metabasites, occurring as basaltic dykes and gabbro stocks. Many of the dykes contain abundant ilmenite, and are ferrobasaltic in composition. The most narrow, less than one metre thick, of these dykes are frequently thoroughly schistose and chloritic, whereas thicker dykes and small bodies can have amphibolic or even metagabbroic core parts (Kontinen 2005).

The presently known sulphide mineralizations in Kylylahti occur as an approximately 1.3 km long trail of several semimassive-massive quartz-sulphide lenses along the steeply dipping eastern margin of the massif. The sulphide mineralizations are found in association to an especially thick zone of Outokumpu assemblage carbonate-skarn-qartz rocks and black schists. The sulphide mineralisation are found in two types. (1) long, narrow semi-massive to massive sulphide layers and lenses, located at the interfaces between the carbonate-skarn-quartz rocks and black schists. These mineralizations themselves are quartz rocks, and have usually very sharp borders without any gradation either to carbonate-skarn-quartz rocks or black schists. (2)Skarn-hosted sulphide disseminations in the carbonate-skarn-quartz rock zones. This disseminated ore includes minor sulphide veinlets, blotches and semimassive sulphide lenses that parallel the strike of the primary ore lenses. The upmost part of the deposit is called Wallaby, and it consists of several ore lenses, 5 m thick and 50 m high at most. Lower part is called Wombat, and has a maximum thickness of 30 m and height of 170 m. The area where upper and lower parts overlap is called Gap

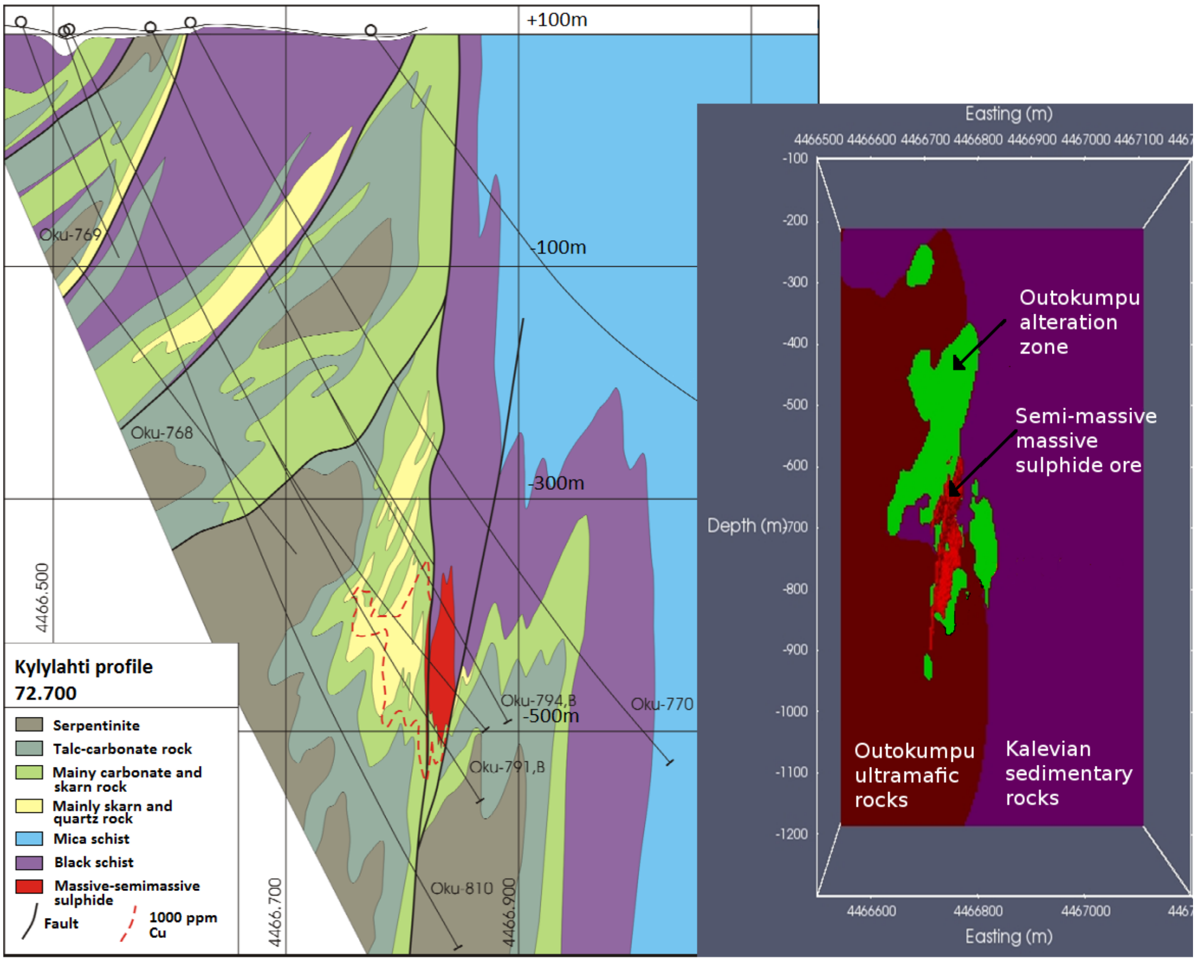

Figure 5: Cross-section of the Kylylahti mineralization, sketching the geological setting of the deep massive-semimassive sulphide lens. The 1000pm Cu isopleth constrains a maximum distribution of the ore-grade Co-Cu sulphide disseminations flanking the massive-semimassive sulphides. Map location of the profile is shown in Figure 4. On the right, a simplified geological model of Kylylahti deposit, used as a starting point for COGITO-MIN seismic data analysis. Modified from Kontinen et al. (2006) and Riedel et al. (2017b).

(Figure 4). A possible down or southward continuation of the deposit is under investigation (Kontinen 2005, Peltonen et al. 2008).

The Kylylahti ore, a Cu-Co-Zn-Ni-Ag-Au sulphide mineralization, contains pyrite, pyrrhotite, chalcopyrite ± sphalerite, sometimes cobaltite and pentlandite. The sulphide mineralizations are mainly pyrite dominated, but pyrrhotite takes more place towards the depth. The main gangue mineral is quartz. Other gangues include calcite and tremolite. Mica, rutile, sphene and tucholite are found as trace minerals. Five types of sulphide mineralizations can be specified, of which types one and two are the most common. 1) The banded pyrite type, which is low in Cu, Co and Zn and its banded appearance is due to alteration of pyrite- and quartz-dominated bands. 2) Blebby pyrite type has large, several cm-size pyrite blebs consisting of pyrite+chalcopyrite±sphalerite±pyrrhotite in a matrix of coarse-grained granoblastic quartz. 3) Pyrrhotitic type, which has been interpreted to represent metamorphosed equivalents of the pyritic mineralizations. 4) Pyrrhotite-magnetite type can be found from the interiors of the deep ore Wombat, in a five to ten metres thick layer. The minaralization contains two to twenty centimetres thick layers, rich in granular magnetite (up to 50 vol.%) in an otherwise mainly pyrrhotitic mineralization. 5) The lowermost edge of Wombat contains a small volume of particularly Co-As-rich mineralization, with a gangue dominantly consisting of carbonate with some tremolite, quartz, rutile and pale-brown biotite. This As-rich minaralization is pyrrhotite-dominant with variable chalcopyrite, sphalerite, cobaltite and pyrite (Peltonen et al. 2008).

Disseminated sulphide mineralizations in Kylylahti are mostly strongly pyrrhotite dominant, with 10-20 vol.% of pyrite and some pyrite rich zones. Magnetite rich parts occur within the more voluminous deep disseminations. Disseminated mineralizations are largely confined to massive tremolite-carbonate material, secondary to and replacing older carbonate-skarn-quartz rocks. The replacive material seems to contain some titanite, rutile, zircon, plagioclase, biotite-phlogopite and apatite, all by definition lacking in normal Outokumpu-type carbonate-skarn-quartz rocks. Similar, replacive, just higher grade metamorphic, diopside skarns can be found from Outokumpu, Vuonos and Luikonlahti (Kontinen 2005).

The bedrock surrounding and enclosing the Kylylahti massif consists of mica schists, with sulphidic black schist intercalations. The immediate wall rock of the ore lenses in Kylylahti is usually black schists, with variable skarn replacement and quartz-sulphide veining. In the east, black schists grade abruptly to mica schists. Kylylahti black schists are heavily sulphidic. Pyrite is the main sulphide phase with variable pyrrhotite, minor sphalerite and some chalcopyrite. In the black schist layers in direct contact to serpentinites, pyrrhotite is dominant. The most sulphidic black schists are separated to their own rock type, sulphide-rich black schist. These rocks are rich in copper, zinc and nickel, but low in cobalt, in contrast to the adjacent copper-cobalt mineralizations (Kontinen 2005).

2.6 Areal geophysics

Outokumpu district has been a target area for exploration for a long time, and therefore vast amount of geophysical data are available. A summary of the existing geophysical data was conducted during the GEOMEX project, and reported by Ruotoistenmäki and Tervo (2006). Since the GEOMEX project ended already more than a decade ago, more data has been accumulated. Petrophysical data has been gathered especially by mining companies acting in the area, Mondo Minerals B.V. in Horsmanaho and Boliden in Kylylahti. Another large set of petrophysical data comes from the Outokumpu Deep Drill Hole, a 2516 m deep research bore hole drilled in years 2004 and 2005. The entire project took place between the years 2003 and 2010. Its results were reported in Kukkonen (2011) and in several other published articles. Petrophysical properties of the Outokumpu Deep Drill Hole core, based on laboratory measurements of the core samples, were described by Airo et al. (2011) and Elbra et al. (2011). Heinonen et al. (2011) reported petrophysical results aquired by downhole logging the Outokumpu Deep Drill Hole.

During the Finnish Reflection Experiment(FIRE, years 2001-2006), the deep crustal structures of Finland were imaged (Kukkonen and Lahtinen 2006). Deep structures of the Outokumpu district were researched analyzing FIRE lines FIRE-3, OKU-1, OKU-2 and OKU-3. FIRE-3 was aimed to study crustal-scale structures, whereas OKU-lines had higher resolution and were used to get a detailed view of the uppermost few kilometres of the crust. Results were used to position the Outokumpu Deep Drill Hole to enclose one of the strong deep reflectors seen in the data, expected to be a previously unknown occurrence of Outokumpu assemblage rocks. Integration of FIRE data and Deep Drill Hole data was reported by Heinonen et al. (2011). After FIRE project, High Resolution Reflection Seismics for Ore Exploration (HIRE, years 2007-2010) was launched. It's aims were in introducing reflection surveys as an exploration tool for the Precambrian crystalline bedrock of Finland and adding on information of the study areas. One of the study areas was Outokumpu district. HIRE project was discribed by Kukkonen et al. (2011) and the results covering the Outokumpu district by Kukkonen et al. (2012b).

The regional petrophysical database of GTK, recently described by Airo and Säävuori (2013), currently contains petrophysical data on more than 130 000 outcrop samples systematically collected over the whole Finland. The database entries contain location coordinates, rock class attributes and determinations of density, magnetic susceptibility and intensity of remanent magnetization. Petrophysical sampling methods, data aquisition and measurements equipment are the same as those used for samples measured for this work. The regional database samples from the Outokumpu district, more than 600 samples, do not contain Outokumpu assemblage rocks. Instead they are ideal for characterizing the physical properties of regional lithological units. Summary of the results for the Outokumpu district can be found in Leväniemi (2016).

Almost the whole of Finland has been covered by high-resolution aerogeophysical surveys, carried out by GTK during 1972-2007. This data includes simultaneously measured magnetic, radiometric and electromagnetic data. The data was recorded by flying at 40 metres height. Distance between the flight lines was 200 metres, and data points at every 50 metres along the lines. Exploration of massive sulphide deposits was one of the main reasons to start systematic airborne geophysical surveys in Finland. The charactericstics of Outokumpu-type sulphide deposits in airborne geophysical data have been described by Airo and Loukola-Ruskeeniemi (2004) and the entire aerogeophysics project in Airo (2005).

Some views on the aerogeophysical data over the Outokumpu district, magnetic anomaly, apparent resistivity and apparent uranium concentration, are shown in Figures 6 and 7. Also the geological map over the area is shown to ease the interpretation. Data gathering and processing has been described by Hautaniemi et al. (2005). Magnetic anomalies due to bedrock, overburden and manmade constructions are revealed by correcting the data by removing the effects of measuring process and time (data is reduced to year 1965.0) and removing the reference field DGRF1965.0.

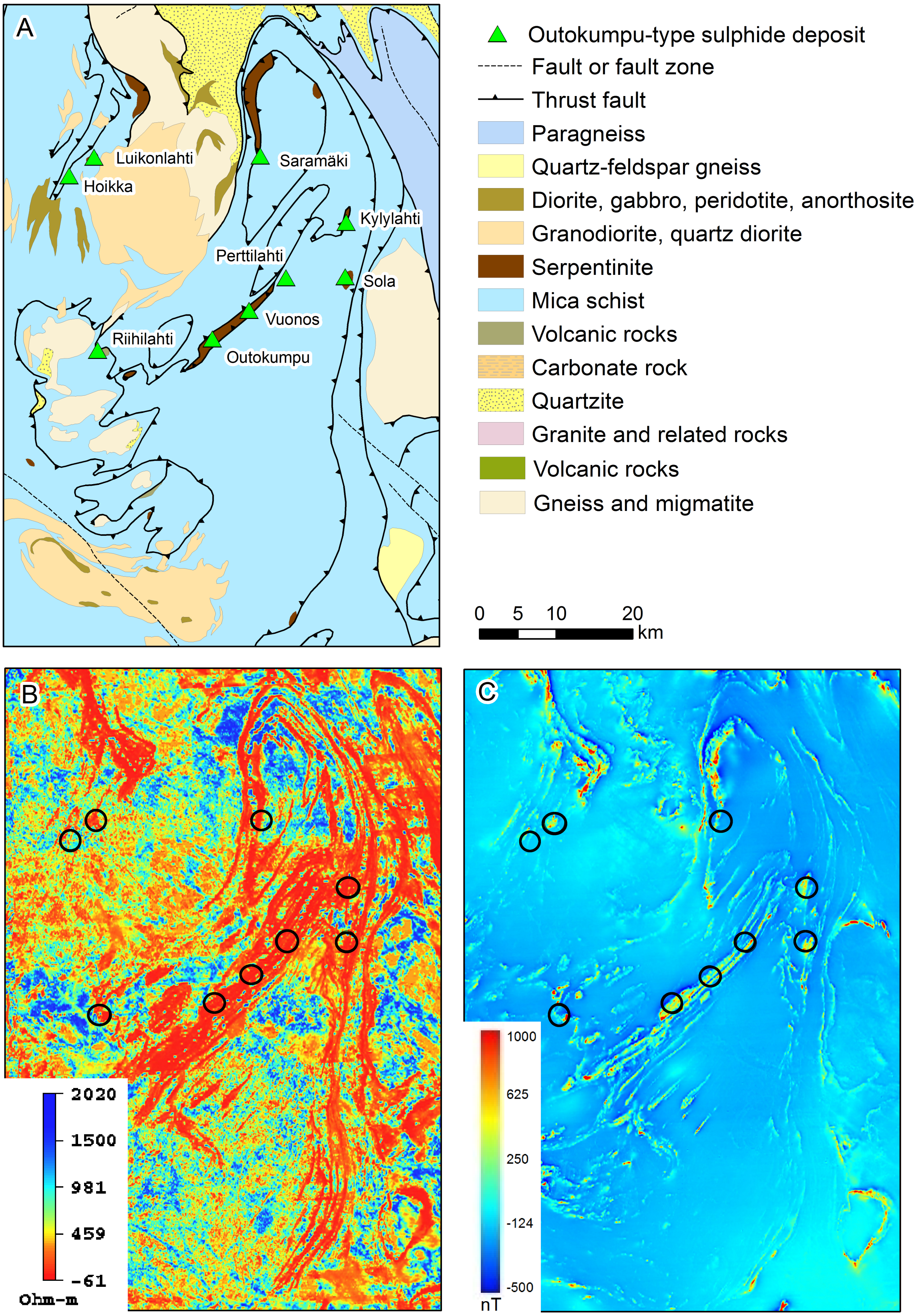

Figure 6: Aerogeophysical data over outokumpu district. A:Geological map of Outokumpu district, B):Apparent resistivity and C:Magnetic anomaly. Outokumpu-type sulphide deposits are marked to the maps. Map modified from Generalized bedrock of Finland ©GTK 2018 and Aeromagnetic and eroelectromagnetic maps of Finland ©GTK 2018

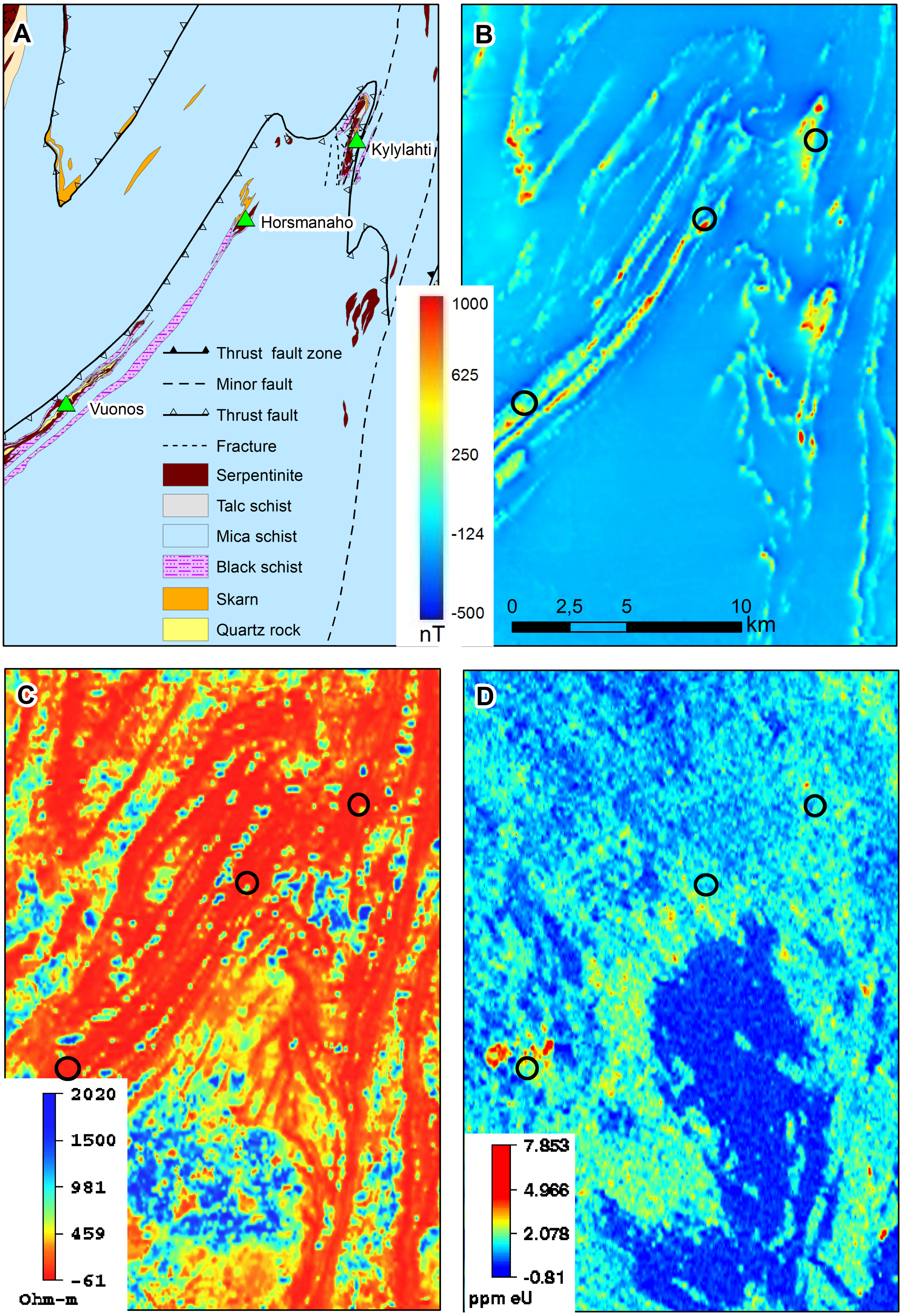

Figure 7: Aerogeophysical data over Kylylahti-Vuonos area. A:Geological map of Kylylahti-Vuonos area, B:Magnetic anomaly, C:Apparent resistivity and D:Apparent equivalent uranium concentration. Vuonos, Kylylahti and Horsmanaho mine sites are marked to the maps. Map modified from Bedrock of Finland ©GTK 2018 and Aeromagnetic, Aeroelectromagnetic and Aeroradiometric uranium maps of Finland ©GTK 2018

Electromagnetic data has been interpreted to apparent resistivity. Primary in-phase and quadrature electromagnetic components can be transformed to apparent resistivity and depth using a half-space model. The map here shows only the apparent resistivity with no depth information. The radiometric data is corrected from the effects related to measurement procedure. The data can then be converted to apparent radioelement concentrations of which apparent equivalent uranium concentration is shown here. Values are in equivalent uranium concentration because the measuring window for uranium catches also radiation caused by bismuth-214, and therefore these two cannot be separated from each other in the results.

From Figures 6C and 7B it is apparent, that Outokumpu-type deposits are associated with strong magnetic anomalies. These anomalies are mainly due to serpentinites and black schists that contain magnetite and monoclinic pyrrhotite, respectively. Within the mineralized area, the magnetic anomaly can be weaker than in the rocks surrounding it, due to pyritization of pyrrhotite (Airo and Loukola-Ruskeeniemi 2004). Geological structures can also be seen on the magnetic map. Accurate structural interpretation needs to be done combining geophysical, petrohysical, geological, and topographic data. Using gradients and various filtering of data is often useful (Airo et al. 2014). Still, even the basic magnetic anomaly (Figures 6C and 7B) depicts how magnetic anomalies align with geological structures. Boundaries of bedrock units, especially shear zones, appear as the areas of weakened magnetization, due to the loss of magnetite. That is a consequence of enhanced fluid circulation (Airo and Loukola-Ruskeeniemi 2004). Weakening of the magnetization can be seen e.g. along the fault line following Vuonos-Kylylahti direction, and turning southwards at north of the Kylylahti serpentinite massif, in Figure 7B.

According to Airo and Loukola-Ruskeeniemi (2004), coincident electromagnetic anomalies characterise the Outokumpu-type deposits, but the mineralized zones themselves often reveal reduced magnetic intensity and electrical conductivity. Serpentinites and black schists both cause magnetic anomalies. Electrically conductive black schists can be separated from poorly conductive serpentinites, when comparing magnetic and electromagnetic maps. In the maps 6B and 7, the high conductivity of black schists is mixed with the conductivity of the fracture and fault areas. In the Kylylahti massif area the highly magnetic northernmost serpentinite tip has a high value of apparent resistivity, whereas the black schists on the western side of the massif have high magnetic anomalies, but low resistivity. The high conductivity of black schists seems to effectively mask the conductivity of the sulphide mineralizations on the resolution of airborne electromagnetic data. In the Outokumpu district the black schists are much more voluminous than the mineralizations.

In the statistical analysis of aeroradiometric data, enhanced equivalent uranium and thorium radiations characterize the sites of the Outokumpu-type deposits. The mineralized zones themselves are depleted in equivalent thorium and potassium radiation, why they have anomalously high uranium to thorium ratios (Airo and Loukola-Ruskeeniemi 2004). These statistical characteristics are not easily found by a visual inspection. In the apparent equivalent uranium concentration map in Figure 7D the most prominent features are high uranium concentrations in Vuonos area. Human activities such as mining districts, landfill, or wastewater ponds appear in general as sources of enhanced radiation (Airo and Loukola-Ruskeeniemi 2004). The big area of negative equivalent uranium values on the southern parts of the map is lake Viinijärvi. Water effectively attenuates the radiation (Airo et al. 2014).

Gravity data was successfully used to reveal the deep-plunging extension of the Outokumpu Belt between Vuonos and Kylylahti already in the 1980's (Ahokas 1984). As a result of decades of intensive exploration, Outokumpu district is well covered by gravity surveys. The data was reviewed and used for potential field modelling by Leväniemi (2016). Two examples in her study were from Kylylahti: a 3-D inversion model constrained by density data, and a synthetic example of the gravity response of the Kylylahti deposit. Leväniemi (2016) concluded that gravity modelling together with reflection seismic data and geological constraints can be used to target reflectors containing Outokumpu assemblage rocks, even to the depths of several kilometers. However, the interpretation of the deep reflectors remains speculative. From the Kylylahti synthetic example it was found out that a body with dimensions and properties similar to the Kylylahti massif can be detected in gravity data from a depth of one kilometre or less.

3 Petrophysical datasets used in this study

In this work a new set of petrophysical data based on the results of COGITO-MIN petrophysical measurements is presented. In the discussion and when choosing the recommended petrophysical parameters for seismic modelling, all relevant available petrophysical data from the study area are used. Other datasets include regional petrophysical database of GTK and petrophysical data from drill cores. This data was recently classified and summarized by Leväniemi (2016). The data selected for her analysis amounted to ca. 38000 density and 26000 magnetic susceptibility values. These summaries have been used directly to avoid repetition.

One big petrophysical dataset from the Outokumpu district that is not included in Leväniemi (2016) is the petrophysical data from the Outokumpu Deep Drill Hole and Core. This data will be used as a reference as it is the only source of measured seismic velocities for the Outokumpu assemblage rocks, other than the measurements done for this work. It is also otherwise an extensive petrophysical dataset.

Boliden Kylylahti is a partner in COGITO-MIN-project, and therefore the entire Kylylahti geophysical database (Boliden 2016b) was available for this study.

239 samples in total were measured for COGITO-MIN-project. The majority of the samples, 209, were provided by Boliden Kylylahti. These are called Boliden samples. Another sample set containing 30 ore samples from several sites in the Outokumpu district, was provided by GTK geologist Asko Kontinen and is called ASKO sample set. ASKO samples together with the Boliden sample set ore samples (massive or semi-massive ore) are called the ore sample set. The origin of the samples is listed in Table 3.

Table 3: Sample sets

Sample set	Number of samples	Description
CG_MIN_901_1-138	137	From the borehole KU-901
CG_MIN_900_1-24	24	From the borehole KU-900
CG_MIN_1_7-10	4	Hand specimen from the excavated ore pile 650jp2
CG_MIN_2_1-3	3	Hand specimen from the excavated ore pile ROMPAD
CG_MIN_999_1-15b	16	From the borehole OKU-999
CG_MIN_1000_1-9	10	From the borehole OKU-1000
CG_MIN_1002_1a-9	10	From the borehole OKU-1002
CG_MIN_1003B_1-5	5	From the borehole OKU-1003B
ASKO	30	Hand specimen from the collection of Asko Kontinen (GTK, Kuopio). Ore samples from several mining sites in the Outokumpu district.

Most of the samples were from 6 drill holes, covering lithological units from surface to approximately 850 metres depth in Kylylahti. These samples had been classified according to GEOMEX conventions (rock type and lithology) by Boliden geologists. Seven ore samples were from the piles of excavated ore in Kylylahti. Those were needed to provide a few ore samples big enough to be cutted to inch by inch cubes, for future pressurized P-wave velocity measurements. These samples had no coordinates but they were classified to be massive or semi-massive ore samples from Kylylahti. Borehole core samples containing significant amounts of sulphide mineralizations were not big enough to be used for pressurized measurements, because a half of the core had been grinded and used for compositional determinations.

ASKO set samples were hand specimen from different mining sites in the Outokumpu district and Kivisalmi boulder. These had no exact coordinates. ASKO samples are mostly ore samples. A few exceptions are two samples of sulphide-rich black schist, and one sample of quartz rock, that could be classified as disseminated ore. ASKO samples were classified by Asko Kontinen.

The rock types of Boliden samples, as well as the number of samples of each type, and abbreviations used for them can be found in Table 4. Abbreviations, sampling sites and classifications for ore samples are in Table 5. It should be noted that Boliden set ore samples are classified as massive or semi-massive. ASKO set ore samples are just ore samples. Disseminated ore has not been separated to its own rock type, but disseminated ore samples are classified after their host rock classes. They belong mainly to quartz rocks or to carbonate rocks.

Table 4: Rock types of Boliden samples and their abbreviations

Abbreviation	Number of samples	Rock type
MS	13	Massive ore
SMS	21	Semi-massive ore
SP	37	Serpentinite
SKA	7	Skarn
TRESKA	22	Tremolite Skarn
QTZR	43	Quartz rock
CRBR	13	Carbonate rock (Outokumpu assemblage)
SS	11	Soap stone
CHLS	10	Chlorite schist
BS	3	Black schist
SULBS	13	Sulphide-rich black schist
MCAS	6	Mica schist
TRECS	6	Tremolitic calc-silicate rock or tremolite schist
CRB	2	Carbonate rock (Kalevian rock)

Table 5: Classification of ore samples and their abbreviations.

Abbreviation	Number of samples	Description
MS	13	Kylylahti (Boliden samples), Massive ore
SMS	21	Kylylahti (Boliden samples),Semi-massive ore
LLAS	8	Luikonlahti, Asuntotalo ore
LLKU	4	Luikonlahti, Kuparivuori ore
LL-QTZR	1	Luikonlahti, Quartz
LL-SULBS	1	Luikonlahti, Sulphide-rich black schist
VU	6	Vuonos, Vuonos ore
OUMO	2	Outokumpu, Mökkivaara ore
OU	5	Outokumpu, Outokumpu ore
KI	1	Kivisalmi boulder, Outokumpu ore
KY	1	Kylylahti, Kylylahti ore
KY-SULBS	1	Kylylahti, Sulphide-rich black schist

4 Measurement procedure and petrophysical parameters determined

Petrophysical laboratory measurements were conducted during the spring and the summer of 2016 in the petrophysical laboratory of the Geological Survey of Finland Espoo office. The parameters determined in the measurements were density, seismic P-wave velocity, porosity, magnetic susceptibility, intensity of remanent magnetization, inductive resistivity, galvanic resistivity and chargeability. Additional parameters calculated from the measurements were seismic impedance, Königsberger (Q) ratio and induced polarization (IP) estimates.

Geological Survey of Finland (GTK) petrophysical laboratory in Otaniemi, Espoo possesses a broad selection of geophysical research equipment. The laboratory has a long history in measuring rock samples and developing measurement devices. It was founded by Mauno Puranen in the 1960s, and further developed by Risto Puranen. All the measurement devices used for this work have been designed and built at the GTK. The devices are shortly described below with an introduction to each parameter determined. Accuracy and precision of each type of measurements (Table 6) are also discussed.

Table 6: Accuracy of measurements

Measurement	Measuring range	Accuracy
Density ρ	1500-20000 kg/m^3	Relative standard error 0.1 % Systematic error < 1 %
P-wave velocity V_P	not specified	10 m/s
Porosity PE	0-100 %	Standard error 0.01 %
Susceptibility k	$10^{-2} - 5 \cdot 10^{6}$ $[10^{-6}]$ SI	Standard error <200 µSI, when $k < 1000$ µSI Relative standard error even more than 10 %, when $k > 1000$ µSI
Remanent magnetization J	$10^{-3} - 10^{3}$ A/m	Standard error 10 mA/m
Inductive resistivity R	$10^{-2} - 10^{-7}$ Ωm	Relative standard error <10 %, when $k < 0.2$ SI
Galvanic apparent resistivity R	0.01-100 000 Ωm	Relative standard error <10 %

4.1 Sample preparation

Before the measurements samples needed to be mechanically prepared. Original samples were mostly pieces of drill core, with a diameter of 5 cm. All the drill core samples with significant amounts of sulphide mineralizations were halves of the original drill core. The other half had been grinded and used for compositional analysis. With drill core samples, cutting them to a proper length and smoothening the ends was enough. ASKO samples and some Boliden ore samples were hand specimen. From those a core sample was first drilled out, with a diameter of 4cm. Then the sample was cutted and its ends smoothened. A few hand specimen were too small to be processed. They were left in their original shape, except for levelling and smoothening two opposite ends.

The target length for diameter 5 cm full drill core samples was 7.5 cm, and for diameter 5 cm half drill core samples 6.5 cm. For diameter 4 cm drill cores the target length was 6cm. The optimal length of samples was determined by requirements of inductive resistivity measurements. The height to diameter ratio should be 1.5 at most (R. Puranen et al. 1993a). For the other measurements, the samples should preferably be bigger than 50 cm^3 (S. Vuoriainen 2016, pers.comm.). The target lengths were achieved for most of the samples. The average length for ⌀5 cm full drill core samples was 7.2 cm, and for half drill core samples 6.3 cm. For hand specimen the average length became only 5.1 cm. The measurement devices for magnetic susceptibility, remanent magnetization and inductive resistivity mark a warning in the results when a sample is smaller than 50 cm^3. That was the case for 12 samples. All these result were still taken into account when calculating averages for the rock types. The unaccuracy caused by the small sample size was considered to be less important, than the need to have as many samples as possible for each rock type. The values for these small samples were within the variation of the values for bigger samples of the same rock type. Magnetic and electrical properties are typically highly variable and may show anisotropic properties, even within the same rock type and unit (e.g. M. Puranen and

R. Puranen 1977) so small inaccuracies are masked in the normal variation of the values.

4.2 Measurement procedure

After preparation, the samples were first measured with the GTK basic petrophysical measurement set protocol. The set, consisting of density, susceptibility and intensity of remanent magnetization measurements, provides a rapid system for measuring a large number of samples with sufficient accuracy for the ordinary needs of geophysical interpretation and rock type classification. The basic set of measurements also gives inductive resistivity as a byproduct of the susceptibility measurement. This set of measurements has been made for example for the outcrop samples of the GTK regional petrophysical database, containing more than 130 000 entries.

In addition to the basic set of measurements, some additional measurements were done. These were determination of seismic P-wave velocity, specific resistivity by galvanic method and porosity. After the basic measurements, the samples were put in water for ten days before the first P-wave velocity measurement. After that the P-wave velocity was measured twice more in ten days intervals. Repeated P-wave velocity measurements were made in order to see the effect of water saturation time. After the last P-wave velocity measurements (when the samples had been in water for 30 days), specific resistivities were determined by galvanic method. Then the densities of the samples were determined for the second time for porosity measurement. After density determination, the samples were dried in an oven in about 110°C for 3 days. Then they were weighed once more to determine their porosity.

Some parameters helping the interpretation of results were calculated from the measured ones. These were seismic impedance, Q ratio IP estimates. All the parameters, measured or calculated, are shortly described in the following.

4.3 Density

Density ($\rho, [kg/m^3]$) is defined as the quotient of the mass m and the volume V of a material:

$$\rho = \frac{m}{V}. \tag{4.3.1}$$

In this study we discuss only the rock bulk density: the mean density of a considered rock volume, including all the different minerals, pores and pore fluids (Schön 2015).

The density of a rock depends on its mineral composition and porosity. With density data we can e.g. estimate rock types and minerals, their sulphide abundances and their degree of alteration (Emerson 1990). In Finnish chrystalline rocks, porosity is usually very low and thus has very small effect on density.

In chrystalline rocks the density increases with higher content of mafic minerals or iron. Denser rocks have typically higher P-wave velocities (e.g. Airo and Säävuori 2013, Schön 2015). Density is important petrophysical parameter in many geophysical methods, especially in gravity and seismic methods. Gravity methods are used to solve the underground density distribution of the survey area. In reflection seismics, the parameter defining the seismic properties of a rock unit, seismic impedance, is a product of density and P-wave velocity of that rock unit (e.g. Reynolds 2011, and the references therein).

At the GTK, rock bulk density is determined by weighing samples both in air and suspended in water. This method is based on Archimedes' priciple. The density is calculated by the equation:

$$\rho = \frac{m_a \cdot \rho_w}{m_a - m_w}, \tag{4.3.2}$$

where m_a is the mass weighed in the air, m_w the mass measured when the sample is suspended in water and ρ_w is the density of water (Kivekäs 1993, Schön 2015). The relative standard error of repeated density determinations in the GTK geophysics laboratory is less than 0.01%. A small underestimation of density of samples is caused by their porosity, but for chrystalline rocks with low porosity, like the samples of this work, the systematic error is generally below 1 % (Kivekäs 1996).

4.4 Seismic P-wave velocity

P-wave velocity, ($V_P, [m/s]$), is the velocity of compressional wave in an elastic material. An ideally elastic material responses to an applied stress according to Hooke's law: the strain is proportional to the stress causing it, and that proportionality can be described with an elasticity tensor. Crustal rocks are mostly elastic materials, but they cannot be considered as ideally elastic. Their reaction to stress depends also on the velocity and the degree of deformation. Although there are deviations from Hooke's law, it is good enough an approximation when forces and deformations are small, and it can be used when evaluating propagation of seismic waves in bedrock. P-wave velocity can be expressed as a function of density and elastic constants:

$$V_P = \sqrt{\frac{K + \frac{4}{3}\mu}{\rho}}, \tag{4.4.1}$$

where K and μ are elastic constants, bulk and shear moduli, respectively, and ρ is the density (Schön 2015). In this work, the P-wave velocity is measured directly.

The elastic properties of rocks, also the velocity of elastic waves, are primarily controlled by the elastic properties of rock forming minerals, their fractional volume and how they are bonded together. The elastic properties of minerals are controlled by their chemical elements and the bonding and internal structure between the elements. Almost all rocks have also pore fluids that have different kind of elastic properties from the host rocks (Schön 2015). The measurements of the elastic wave velocities are important when interpreting seismic exploration results. The results other than in-situ at the survey site or at laboratory from the survey site rock samples, are not necessarily a reliable reference.

In general, denser rocks have higher P-wave velocities, but there are exceptions. What equation 4.4.1 suggests, is that the higher the density, the lower the P-wave velocity. However, denser minerals are usually also less compressible (higher bulk modulus) and more rigid (higher shear modulus), why they have higher P-wave velocities (Eaton et al. 2003). Increasing porosity is related to lower velocities. Increasing pressure leads to higher seismic wave velocities. The change from higher porosity and lower velocity to lower porosity and higher velocity is the most notable on the low pressure range, in the upper parts of the crust. The effect in shallower depths is due to the closure of micro cracks and pores in the rocks. When going deeper, to higher pressures, increase in the velocities is slower, and it is due to the change of the elastic properties of the minerals of rocks (Schön 2015). In the Outokumpu region, the pressure limit where fast and non-linear rise of the seismic velocities ceases, has been measured to be around 200 MPa, corresponding to depths shallower than 10km (Kern and Mengel 2011). Growing temperature lowers seismic wave velocities. Up to about 100-150 °C the effect is very small, and it becomes important in the higher teperature ranges (Schön 2015).

Seismic data interpretation nowadays may include seismic forward modelling. Although P-wave velocities determined in laboratory do not directly represent the in-situ values on the survey site, they do give solid basis for determining parameters used in models. In addition to seismic velocities, also density values are needed for seismic forward modelling (e.g. Reynolds 2011).

In this work, P-wave velocity was measured using an ultrasonic transducer. In the device an electric pulse with a high frequency of about 1 MHz is generated with a pulse generator, and transmitted through a submerged sample. The transit time is measured with a pulse counter. The velocity of the pulse in the sample can be calculated, when the lenght of the sample is known (R. Puranen and Sulkanen 1985). The measurements were made at room pressure and temperature. Samples were submerged in water to provide good contact between the transducer and the sample. The length of a sample was determined by measuring the sample three times with a digital caliper, every time from a different spot on the cylinder ends. The length used in velocity determination was the average of these three values.

The accuracy of the measured P-wave velocity is approximately 10 m/s. The length of water immersion time before P-wave velocity measurement has a considerable effect on the result. A longer immersion time increases the P-wave velocity (Airo et al. 2011). For that reason the samples were soaked for 30 days and the P-wave velocity was measured every ten days. A weighted average of the measured velocity ($\frac{1}{6}$ weight for the first, $\frac{2}{6}$ for the second and $\frac{3}{6}$ for the third mesurement) was calculated. On average the velocities increased notably from the first to the third measurement, but for individual samples it could be even on the contrary, thus a weighed average was estimated to be a representative value for the real velocity, instead of using the results from the last measurements.

4.5 Seismic impedance

The acoustic impedance ($I, [MPa \cdot s \cdot m^{-1}]$) of an elastic medium is the ratio of the stress to the particle velocity, and is given by

$$I = \rho \cdot V, \tag{4.5.1}$$

where ρ is the density and V is the wave propagation velocity (Mavko et al. 2009). When there is a difference on seismic impedances of two rock bodies, a part of the wave energy is reflected from the interface. The reflection coefficient R_{12} between two isotropic, linear, elastic media for vertical incidence is given by:

$$R_{12} = \frac{I_1 - I_2}{I_1 + I_2} \tag{4.5.2}$$

where I_1 and I_2 are the seismic impedances of these two media (Mavko et al. 2009). Two different rock units are usually not linear nor isotropic medias and waves rarely vertical to the contact, but this scenario can still be used as an approximation. A reflection coefficient of 0.06 is considered to be high enough for a detectable reflection (Salisbury et al. 1996). With reflection coefficient 0,06, the interface reflects 6% of the incident energy. Reflection coefficients can also be negative. In that case polarity of the reflected wave reverses (e.g. Salisbury et al. 1996, Eaton et al. 2003). Seismic impedances are needed when considering which lithological contacts are visible on seismic reflection data, and which remain invisible.

4.6 Porosity

Porosity is the fraction of the entire volume of a rock that is not occupied by solid constituents. Effective porosity($PE, [\%]$) is the part of porosity that is available for free fluids. In petrophysical studies, effective porosity is more useful rock property than total porosity. Measuring total porosity would require the samples to be grinded, which would be expensive and time-consuming. Effective porosity also has more influence on the other properties of rocks, such as permeability and electrical conductivity, than the total porosity (Kivekäs 1994).

Porosity of a rock is a result of various geological, physical and chemical processes. Higher porosity means typically lower density, specific resistivity, and seismic wave velocity. Porous rocks are often also mechanically weaker (Schön 2015). Porosity of rocks is therefore an interesting parameter within geophysical surveys and in rock engineering, for example in mine planning. Precambrian rocks have generally very low porosities. Weathered surface rocks, large-grained, highly altered, and fractured rocks are exceptions (Airo and Säävuori 2013).

At the GTK, the effective porosity is determined by water saturation method. A water saturated sample is weighed, dried in an oven at 110°C for three days and then weighed again. Effective porosity can be calculated from:

$$PE = 100 \cdot \frac{(M_W - M_D)/\rho_w}{V} \tag{4.6.1}$$

where M_W and M_D are water saturated and dry weights of the sample, respectively. ρ_w is the density of water, and V is the sample total volume. Three days is usually assumed to be sufficient length for saturation time (Kivekäs and R. Puranen 1995). In this study the samples were soaked for 30 days, due to P-wave velocity measurements. The accuracy of porosity measurements at the GTK petropysics laboratory is better than 0.01 % for samples bigger than 20 cm^3 (Kivekäs 1993). All the samples used in this work were over that size.

4.7 Magnetic susceptibility

Magnetic susceptibility ($k, [\mu SI]$) is a measure of the ability of a rock to become magnetized when exposed to an external magnetic field. The magnetic susceptibility (later susceptibility) is defined as the relation between the induced magnetization M and the applied magnetic field H:

$$M = k \cdot H. \tag{4.7.1}$$

Based on their magnetic properties, materials can be divided into three groups: 1) diamagnetic, 2) paramagnetic, and 3) ferro-, ferri-, and antiferromagnetic materials. In diamagnetic minerals the external magnetic field causes a weak opposing field, why their susceptibility is small and negative. For example quartz and plagioclase are diamagnetic minerals. Diamagnetic susceptibility is independent of temperature (Schön 2015).

Paramagnetic materials also produce a weak magnetic field in the presence of an external field. The produced magnetization is in the same direction than the external field causing it. The susceptibility of paramagnetic minerals is positive. Some examples of paramagnetic minerals are olivine and pyroxenes.

Susceptibility of paramagnetic materials decreases with increasing temperature (Schön 2015). In Finland, about 75 % of rocks fall into a population dominated by paramagnetic minerals (Airo and Säävuori 2013).

Paramagnetic and diamagnetic materials produce a magnetic field only in the presence of an external field. In ferromagnetic, ferrimagnetic and antiferromagnetic materials the magnetic interactions inside the atoms are so strong, that these materials can carry magnetization even in the absence of an external field. Important groups of ferro- and ferrimagnetic minerals are iron, iron-titaniums oxides and iron sulphides. In ferromagnetic minerals magnetic domains, small volume elements carrying magnetization even without external magnetic field, are all oriented in the same direction and thus create a macroscopic external magnetic moment. Iron is ferromagnetic. In ferrimagnetic minerals the magnetic domains have antiparallel orientation, but different magnitudes, thus resulting an external moment. Magnetite and pyrrhotite are ferrimagnetic minerals. In antiferromagnetic minerals magnetic domains are antiparallel and equal in magnitude, resulting in a zero external moment. In some cases, the domains can be slightly canted, resulting in a weak external moment. Hematite is a canted antiferromagnetic mineral (Schön 2015).

Ferro-, ferri- and canted antiferromagnetic minerals (later magnetic minerals), have higher positive susceptibility than paramagnetic minerals. Magnetic minerals can also carry remanent magnetization. Susceptibility of magnetic minerals is dependent on temperature. Over a certain temperature range, magnetic materials have paramagnetic properties. Their susceptibility is also dependent on magnetic history of the material, because magnetic domains stay in their preset orientation unless they are reset in high temperatures or in high external magnetic field (Schön 2015). Magnetic minerals dominate the magnetic susceptibility of a rock, even at small concentration. Magnetic susceptibility mainly reflects iron and magnetite content in a rock. In the GTK databases, the limit between paramagnetic and magnetic rocks has been set to 2000 µSI (Airo and Säävuori 2013).

Susceptibility is the property of rocks that causes anomalies mapped in magnetic surveys. Remanent magnetization can make interpretation of magnetic data difficult in some cases. When different lithological units have different magnetic properties, they can be distinguished from each other. Also tectonic structures, like shear zones and faults, can be traced from the magnetic data (e.g. Reynolds 2011).

At the GTK petrophysics laboratoratory susceptibility is measured with a low frequency (1025 Hz) AC bridge, composed of two coils and two resistors. When a sample is inserted in the measuring coil, the signal voltage of the bridge will be changed, and the change is proportional to the magnetic susceptibility of the sample. The measuring field parallel to the coil axis roughly corresponds to the present Earth's magnetic field in Finland. Measured susceptibility is corrected for the effects of the shape of the sample (M. Puranen and R. Puranen 1977).

For weakly magnetic (susceptibility<1000 µSI) and nearly isotropic samples, the standard error of repeated measurements is below 200 µSI. For stronger magnetic and anisotropic samples the measurement results can vary by more than 10 %, depending on position and direction of the sample within the measuring coil (R. Puranen et al. 1993b).

4.8 Remanent magnetization

The total magnetization of a rock is composed of induced and remanent magnetization. Rocks that contain magnetic minerals can remain magnetized even when the external field has been removed. The remaining part of the magnetization is called natural remanent magnetization, NRM ($J, [mA/m]$) and it can consist of several different components of magnetization, that have been formed in the history of the rock (e.g. Schön 2015).

Natural remanent magnetization (later remanence) is carried by magnetic minerals and its strength depends on the content, grain size, and shape of magnetic minerals. Remanence is primarily acquired when the rock is formed, but it can be totally or partially overprinted as a result of metamorphic processes (Airo and Säävuori 2013). Remanence can have a different direction and intensity than the induced magnetization. Therefore the remanence should be taken into account when interpreting magnetic data and in magnetic modelling (e.g. Reynolds 2011).

In this work, remanence was measured with a fluxgate magnetometer inside a magnetic shielding. When a sample is placed inside the magnetometer, it causes a change in the magnetic field. The change is proportional to the remanence of the sample in the direction of the fluxgate element. The measuring sensitivity is about $3 \cdot 10^{-3}$ A/m and the standard error 10 mA/m for weakly magnetic samples of a typical size (around 200 cm^3). If the sample is oriented, the intensity and direction of remanence (declination and inclination) is measured. For non-oriented samples, only the total intensity of remanence is given (R. Puranen et al. 1993b, GTK 2015). Samples in this study were non-oriented.

4.9 Königsberger Q ratio

Königsberger Q ratio is the ratio of the intensity of remanent magnetization (M_r) to the intensity of induced magnetization (M_i). It is given by

$$Q = \frac{|\mathbf{M_r}|}{|\mathbf{M_i}|} = \frac{M_r}{k \cdot H} \tag{4.9.1}$$

where k is the susceptibility of the sample and H is the intensity of the external magnetic field (Schön 2015). All Q ratios in this study were calculated using value $41 A/m$ for the external magnetic field, which is approximately the average value of the Earth's magnetic field in Finland (Airo and Säävuori 2013).

Q ratio is mainly used to evaluate the relative proportions of induced and remanent magnetization and to estimate the magnetic mineral type and grain size in a rock sample. In general, Q ratio is only relevant for rock samples containing ferro- or ferrimagnetic minerals, indicated by sufficiently high susceptibilities (limit set to k $>$ 0.002 SI at the GTK) (Airo and Säävuori 2013). Q ratio smaller than one typically indicates that the rock contains coarse-grained magnetite. Q ratio higher than 10 indicates the presence of monoclinic pyrrhotite or fine-grained magnetite (Clark 1997, Airo and Säävuori 2013).

If Q ratios in the survey area are above one, remanence has to be considered when interpreting magnetic anomalies. Especially pyrrhotite-rich rocks can carry a relatively strong remanence, and the direction can be oblique to the present magnetic field of the Earth. Even when the Q ratio is close to unity the remanence can modify the amplitude of measured magnetic anomaly and its ignorance can mislead the interpretation, even if the direction of remanence would be in the same direction as the prevailing magnetic field (Clark 1997). In the GTK petrophysical database, remanence accounts for over 50 % of the total magnetization for most of the measured rock types, and it is especially significant for metamorphic rocks (Airo and Säävuori 2013). That is why it is important to know at least the magnitude of remanence in order to be able to determine whether it should be considered when making magnetic interpretation. Naturally the direction of remanence should also be known for accurate interpretation, but the samples are seldom oriented, and the direction is typically not known.

4.10 Specific resistivity

The electrical properties of any material include electric conductivity and polarizability. Electric conductivity ($\sigma, [S/m]$) is the ability of a material to transport charge. In geophysics its reciprocal, specific electric resistivity ($R, [\Omega m]$) is often used. Dielectric polarization is the charge separation in a material, resulted by application of an external electric field. Dielectric polarization is responsible for the time and frequency dependency of resistivity (Schön 2015). As a result it takes some time for constant current flow to develope when applying a current through a material as well as time for the current to drop to zero when the power source is switched off. In petrophysics this property is often referred as chargeability of a rock. One way to evaluate the frequency dependency are so called induced polarization (IP) estimates. The IP estimates compare the resistivity values aquired using different frequencies of electric current (Telford et al. 1990).

By electric and electro-magnetic methods, areas with contrasting conductivities can be found (e.g. Reynolds 2011). According to their conductivity, rocks and minerals can be divided into conductors, semi-conductors and insulators. In conductors, electrons find free paths to move and thus these materials conduct electric currents. Native metals and graphite fall into this category. Rocks containing sufficient amounts of sulphide mineralizations or graphite are thus conductors (Schön 2015).

Semiconductors are materials with electron energy barriers that are slightly higher than the energy available from thermal activation of electrons at room temperature. At higher temperatures, these materials can be conductors, when electrons get enough energy to move. At lower temperatures semiconductors can transport charges by electrons or holes moving across barriers that are lowered by impurities within the material. Ilmenite and magnetite are semi-conductors (Schön 2015).

Silicate minerals and rocks are insulators having very large energy barriers between atoms so that electrons rarely become charge transport carriers. Very dense or absolutely dry rocks have very high resistivities - except for rocks that contain sulphide mineralizations or graphite. On the other hand, most rocks in the bedrock are saturated with water. In these rocks, porosity is the defining factor of their resistivity. Water or other fluid in the pores can be charge carrier, if pores are connected to each other (Schön 2015).

At the GTK petrophysical laboratory specific resistivities can be measured by the inductive or the galvanic method. These methods are applicable on different ranges of resistivity and are thus supplementary to each other. The inductive method is suitable for highly conductive samples, like sulphide mineralization samples. Galvanic method is good for less conductive rocks. IP estimates are calculated from galvanic resistivity results. Some rock samples can be measured with both methods. Most notably, if a rock sample contains disseminated grains of sulphide mineralizations, it gives low resistivity values by inductive and high resistivity values by galvanic method (R. Puranen et al. 1993a). All COGITO-MIN-samples were measured with both methods, but all samples did not give results by both methods.

At the GTK, by galvanic method, the specific resistivity is measured with a MAFRIP meter. MAFRIP measures specific resistivity of a water saturated sample with a 2-point-system of wet electrodes, and on 3 different electric current frequencies (0.1, 10 and 500 Hz). The resistivity of the sample is calculated by measuring the voltage over the sample. This system of several frequencys allows also for determination of the IP estimates (R. Puranen et al. 1995).

Because resistivity is frequency dependent, the galvanic method gives as many resistivity values as there are frequencies used. Resistivity is lower for higher frequencies. If a sample is not measurable with the 3-frequency protocol, a 2-frequency protocol with only two higher frequencies can be used. For extremely resistive samples, MAFRIP gives only lower limit of resistivity. On the other hand, for very conductive samples (with low resistivity), only the upper limit of resistivity can be measured. MAFRIP has a dry electrode system for measuring well conducting samples, but not even that system can measure resistivities lower than 0.01 Ωm. For these samples, the inductive method should be used. Within resistivity range 0.01-100 000 Ωm, the error of the measurement is below 10 % (R. Puranen et al. 1995).

The inductive specific resistivity of an electrically conducting rock sample can be measured simultaneously with its susceptibility. A conducting, permeable sample placed inside an AC bridge coil induces a secondary voltage in the coil and therefore a signal to the bridge output. The bridge is trimmed so that the in-phase component of the signal is dependent mainly on the susceptibility of the sample, and the out-of-phase component is dominated by the sample's conductivity. For cylindrical samples with susceptibilities lower than 0.2 SI measuring errors are below 10% in the resistivity range 10^{-2}-10^{-7} Ωm (R. Puranen et al. 1993a). Samples with higher resistivities do not give any results when measured by the inductive method.

4.11 Induced polarization, IP estimates

Polarizability of materials causes induced polarization (IP). It is a current-stimulated electrical phenomenon observed as a delayed voltage response in earth materials. IP measures how well material is able to retain electrical charge, it's chargeability (Schön 2015).

Induced polarization can be investigated in time- or frequency domain. The induced polarization generally results in a decrease of resistivity with increasing frequency. One measure of the IP effect is the change in resistivity R measured at two frequencies f_1 and $f_2 (f_1 < f_2)$ of a harmonic alternating current. This parameter is called frequency effect FE:

$$FE = \frac{R(f_1) - R(f_2)}{R(f_1)} \tag{4.11.1}$$

It is a dimensionless parameter and is often given as percent frequency effect $PFE = FE \cdot 100\%$ (Schön 2015).

IP effect is due to electrode or membrane polarization. Electrode polarization occurs when an electrically conducting mineral, for example pyrite, is in contact with ionically conducting pore fluid. Charge carriers, electrons and ions, cannot go through the interfaces between minerals and fluid thus the interface constitutes an electrical impedance. Membrane polarization is caused, when some minerals in rock, usually clay or silicates, create local electric fields into the pores of the rock. These fields partly block the flow of charge carrying ions. This inherent inhomogeneity of charge distribution opposes the current flow. The effect is biggest when balance has been reached, so the resistivity of the system decreases with increasing frequency of electric current applied (Schön 2015). One main application of IP methods is to search for disseminated metallic mineralizations. IP estimates can also be used to evaluate porosity and fracturing of the rocks (R. Puranen et al. 1995, Airo 2015).

At the GTK, frequency-based method is used to investigate chargeability of rock samples that have their specific resistivity measured by galvanic method. IP estimates, FE values for MAFRIP measurement frequencies, are calculated. These IP estimates are:

$$PL = 100 \cdot \frac{R(0.1Hz) - R(10Hz)}{R(0.1Hz)} \tag{4.11.2}$$

$$PT = 100 \cdot \frac{R(0.1Hz) - R(500Hz)}{R(0.1Hz)} \tag{4.11.3}$$

$$IP = 100 \cdot \frac{R(10Hz) - R(500Hz)}{R(10Hz)} \tag{4.11.4}$$

PL and PT estimates were calculated for samples that got resistivity values with the normal 3-frequency protocol. IP values were calculated for all the samples that got resistivity values by galvanic method. The most conductive samples, that only got upper limits of resistivity by galvanic method, could not be given any IP estimates (R. Puranen et al. 1995).

5 Results of the petrophysical laboratory measurements

In this chapter, all the results from laboratory measurements are reported as well as parameters calculated from them. All the parameters measured are plotted as boxplots or as scatter diagrams. In boxplots the average value is plotted as a black point. The box extends from the lower to upper quartile values of the data, with a line at the median. The whiskers extend from the box to show the range, covering 5 and 95 percentiles of the data. Flier points are data points past the ends of the whiskers.

Rock types of the samples or the origin of the ore samples are marked as abbreviations. These abbreviations were presented in Tables 4 and 5. In figures visualizing the results, different rock types are colorcoded according to the recommendations established during the GEOMEX project (J.Juurela 2016, written comm.).

5.1 Density and P-wave velocity

Densities (ρ) and P-wave velocities (V_P) of the Boliden samples are plotted in the figure 8. Average values (Avg) and their standard deviations (std) can be found from Table 7 in numerical form.

As expected, the ore samples have highest density values ranging from a bit over 3000 kg/m^3 to almost 4300 kg/m^3. The average density for massive ore (MS) is 3800 kg/m^3 and for semi-massive ore (SMS)

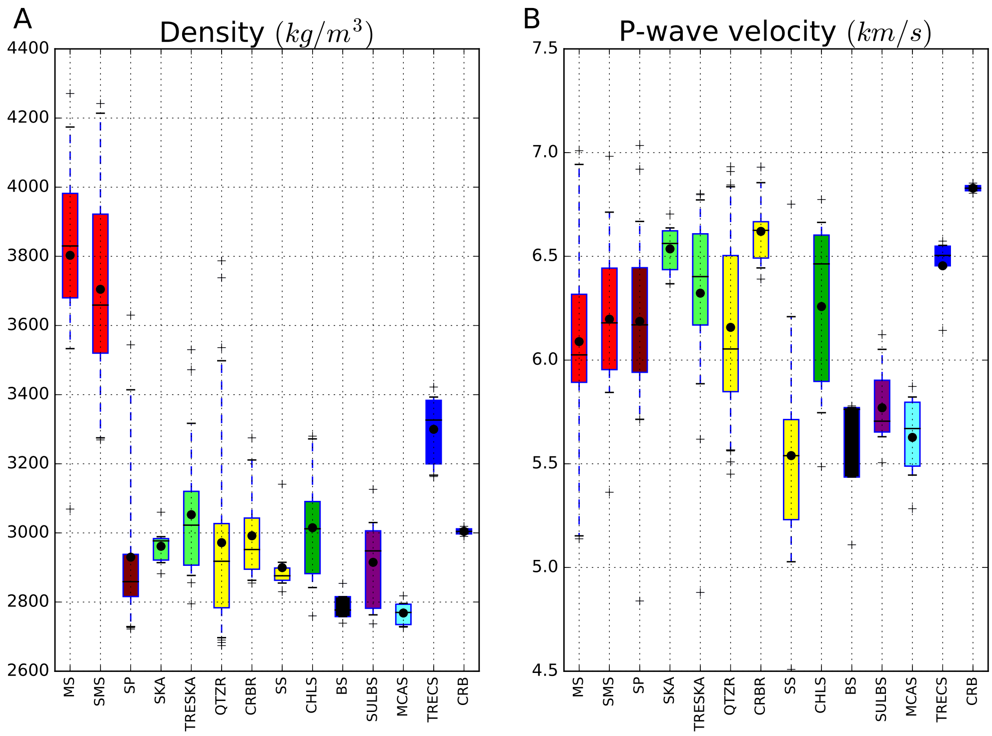

Figure 8: Density and P-wave velocity of the Boliden samples. Rocktype abbreviations are listed in Table 4.

Table 7: Average densities and P-wave velocities of Boliden samples

Lithocode	Rock type	Avg V_P m/s	V_P std m/s	Avg ρ kg/m^3	ρ std kg/m^3	n of samples
MS	Massive ore	6090	550	3803	304	13
SMS	Semi-massive ore	6200	350	3705	316	21
SP	Serpentinite	6190	400	2930	218	37
SKA	Skarn	6540	120	2961	54	7
TRESKA	Tremolite Skarn	6320	430	3053	191	22
QTZR	Quartz rock	6160	420	2972	265	43
CRBR	Carbonate rock - OKU	6620	150	2992	124	13
SS	Soap stone	5540	570	2900	80	11
CHLS	Chlorite schist	6260	430	3015	164	10
BS	Black schist	5550	310	2790	48	3
SULBS	Sulphide-rich black schist	5770	170	2915	122	13
MCAS	Mica schist	5630	210	2768	34	6
TRECS	Tremolitic calc-silicate rock	6450	150	3300	102	6
CRB	Carbonate rock - Kalevian	6830	20	3005	15	2

3700 $kg/^3$. According to Tuomi (2016), the increase in sulphur and iron content correlates with the increase of density for Kylylahti massive and semi-massive ore.

Other rocks than the ore samples have density values mostly between 2800 and 3000 kg/m^3, mica schist having the lowest average density, 2768kg/m^3. As disseminated ore is not defined as its own rock type in this work but disseminated ore samples are classified according to their host rocks, the highest density values in carbonate-skarn-quartz (CRBR, SKA, TRESKA, QTZR) rocks are likely due

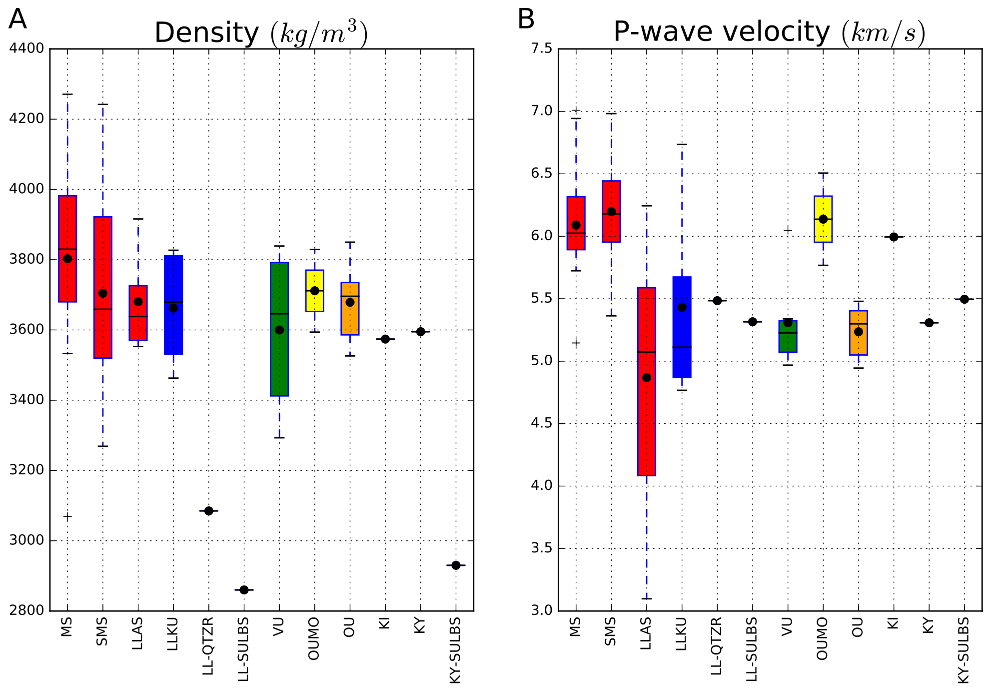

Figure 9: Density and P-wave velocity of the ore samples. Ore-type abbreviations are listed in table 5.

to sulphide disseminations. Also in black schist (BS) and mica schist (MCAS) high densities depend on the enrichment of sulphides in them. The disseminated sulphide content was evident even at the visual examination of these samples. Elevated density of tremolitic calc-silicate rock (TRECS) is due to its magnetite content, which is comfirmed by magnetic results.

In terms of P-wave velocities, almost all the Outokumpu assemblage rocks in Kylylahti are very similar. The averages vary from 6.1 km/s for massive ore (MS) to 6.6 km/s for Outokumpu assemblage carbonate rocks (CRBR), but all the rock types have samples with velocities from both ends of the scale. Massive (MS) and semi-massive (SMS) ore as well as serpentinite (SP) have slightly lower velocities than carbonate rocks (CRBR), skarns (SKA, TRESKA) and chlorite schists (CHLS). The only Outokumpu assemblage rocks with notably low P-wave velocities are soapstones (SS) with an average of 5.5 km/s. Also black schists (BS) and mica schist (MCAS), forming the outer parts or surrounding the Kylylahti massif, have lower velocities, 5.6 km/s on average for black schist and mica schist, and 5.8 km/s for the sulphide-rich black schist.

In Figure 9 densities and P-wave velocities for the ore samples are shown. These densities are mostly between 3600 kg/mm^3 and 3800 kg/mm^3. Massive ore from Kylylahti (MS) shows the highest density. Although the ASKO set samples were not cathegorized to massive and semi-massive samples, based on their density values, it can be assumed that ASKO set samples are mostly of semi-massive type. Sulphide-rich black schist samples (LL-SULBS, KY-SULBS) are well within the range of densities of the sulphide-rich black schist (SULBS) samples of the Boliden sample set. Luikonlahti quartz rock

(LL-QTZR) sample has quite high density, confirming it represents disseminated ore.

P-wave velocities are high for ore samples from Kylylahti (MS, SMS), Outokumpu Mökkivaara (OUMO) and Kivisalmi boulder (KI). This can be readily explained with mineralogical compositions. On these sites, the ore is mostly pyrite dominated whilst in the other sampling sites pyrrhotite is dominant (Kontinen et al. 2006). Pyrite has higher P-wave velocity than pyrrhotite (e.g. Salisbury et al. 1996).

5.2 Seismic impedance

P-wave velocities against densities are plotted for all the Boliden samples (Fig. 10), the averages for the Boliden samples (Fig. 11) and for the ore sample set with the Boliden sample averages (Fig. 12). Seismic impedance isolines are marked to the figures to ease the interpretation. Isolines are spaced in a way that the distance between two lines corresponds to 0.06 reflection coefficient, generally regarded as the difference needed for producing a detectable reflection (e.g. Salisbury et al. 1996). Average seismic impedances as well as reflection coefficients between different geological contacts are also given numerically in Tables 8 and 9.

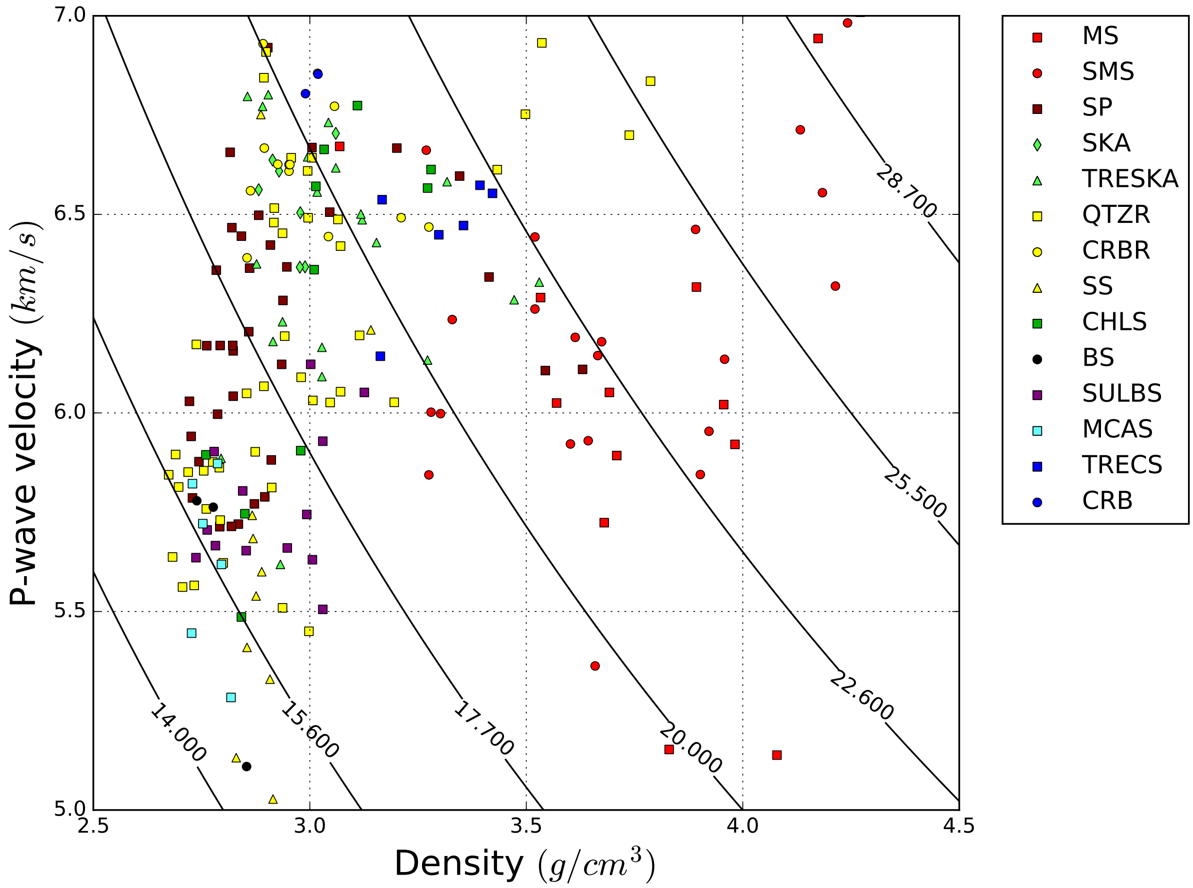

Figure 10: Densities and P-wave velocities of the Boliden samples with seismic impedance isolines. The distance between two lines corresponds to 0.06 reflection coefficient.

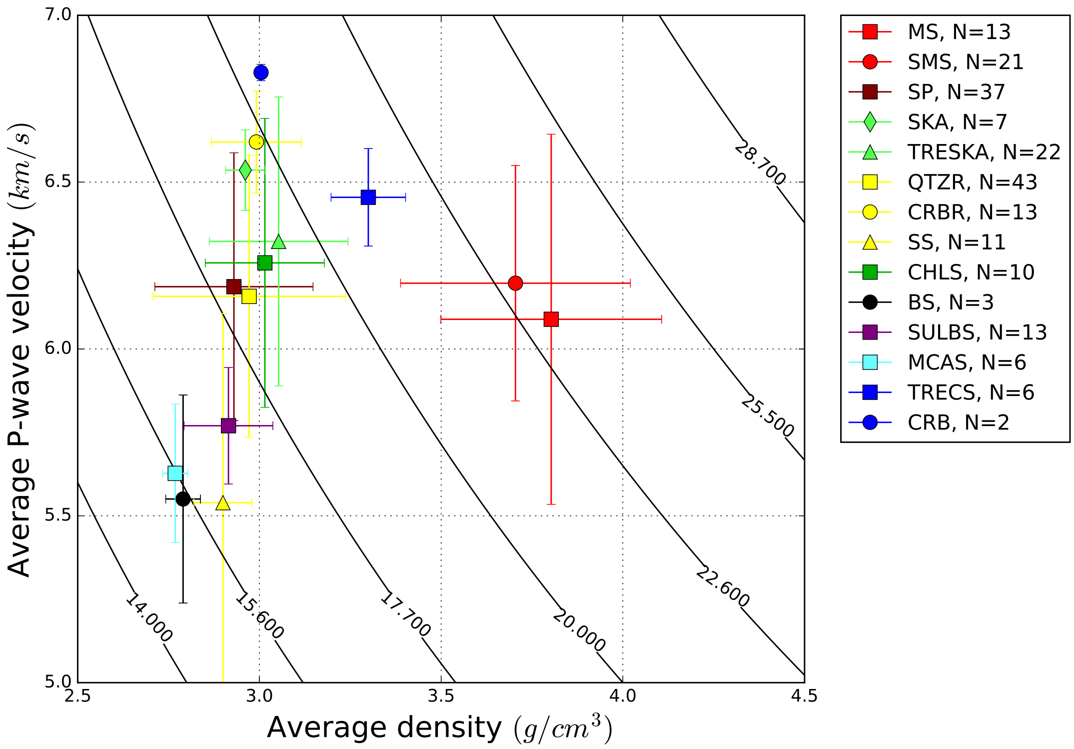

Figure 11: Average densities and P-wave velocities of Boliden sample set rock types with seismic impedance isolines. Standard deviation shown as errorbars. The distance between two seismic impedance isolines corresponds to 0.06 reflection coefficient.

Table 8: Average seismic impedances of Boliden samples $MPa \cdot s \cdot m^-1$

Lithocode	Rock type	Seismic impedance density	Impedance std	n of samples
MS	Massive ore	23.2	2.8	13
SMS	Semi-massive ore	23.0	2.2	21
SP	Serpentinite	18.2	1.9	37
SKA	Skarn	19.2	1.0	7
TRESKA	Tremolite Skarn	19.2	2.0	22
QTZR	Quartz rock	18.4	1.9	43
CRBR	Carbonate rock - OKU	19.7	1.2	13
SS	Soap stone	15.9	2.2	11
CHLS	Chlorite schist	19.0	2.0	10
BS	Black schist	15.6	1.6	3
SULBS	Sulphide-rich black schist	16.9	1.2	13
MCAS	Mica schist	15.5	1.3	6
TRECS	Tremolitic calc-silicate rock or tremolite schist	21.4	1.3	6
CRB	Carbonate rock - Kalevian	20.4	0.4	2

From the figures it can be seen that the ore (MS, SMS) does not have higher P-wave velocity than the rocks surrounding it, but because of its high density, the ore has notably higher seismic impedance than the other rock types. On the other end of the spectrum, mica schists, black schists and soap stones have low seismic impedances. Other Outokumpu assemblage rocks form an intermediate group. Table 9 shows which contacts could produce detectable reflections. The sulphide mineralizations should be detectable against almost any background. Any contact between Outokumpu assemblage rocks, excluding soap stones, and surrounding mica schists and black schists should also be visible, whereas the Outokumpu

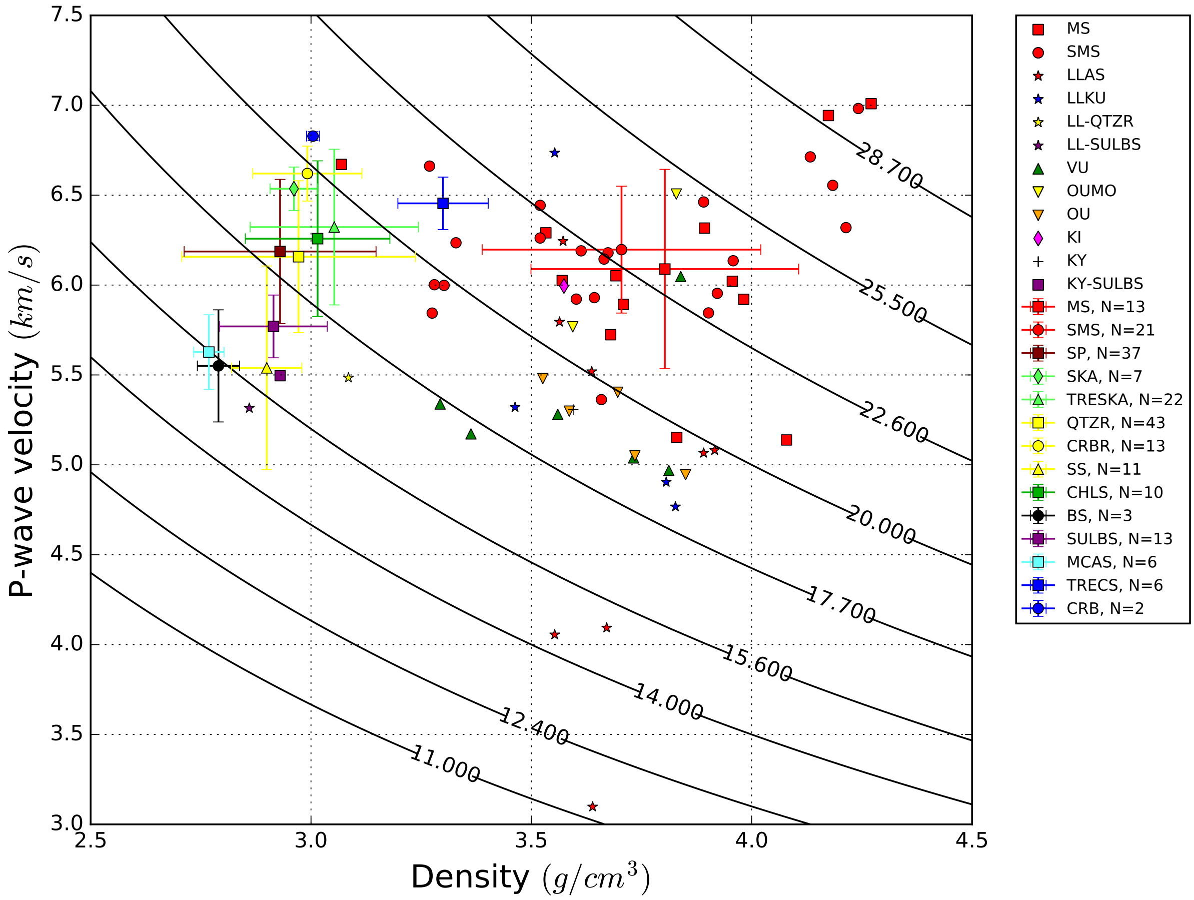

Figure 12: Densities and P-wave velocities of ore samples with seismic impedance isolines. The distance between two lines corresponds to 0.06 reflection coefficient. Also average densities and P-wave velocities of Boliden sample set rock types are shown.

assemblage carbonate-skarn-quartz rocks may not be separable from the serpentinites. Soap stone zones within serpentinite massifs could be detected, if they are voluminous enough.

Boundaries between the ore and other rocks in Kylylahti are usually sharp (Kontinen et al. 2006), thus the ore should produce a detectable reflection if the geometry of the mineralization and the survey allows for it. Kylylahti massif itself should also be a reflector against the schists (MCAS, BS) surrounding it. On the contrary, serpentinites (SP) and carbonate-skarn-quartz-rocks (CRBR, SKA, TRESKA, QTZR) appear to be a seismically homogeneous block, but it should be noted that the range of P-wave velocity values, and thus seismic impedances, for these alteration zone rocks is large. On some locations carbonate-skarn-quartz rocks could produce a reflection in contact with serpentinites. Tremolitic calc-silicate rock (TRECS) and carbonate rock (CRB) have quite high seismic impedances, but due to their smaller volumes they are not expected to produce detectable reflections.

When the seismic impedances of the ore samples from other Outokumpu district mines than Kylylahti are compared to Kylylahti rock type averages (Fig. 12), it appears that the sulphide mineralizations would not necessarily be detectable from the host rocks. On the other hand, also the host rocks may have different seismic properties in other locations. Anyhow these results thus, that it is important to adjust the parameters separately for each survey site, when interpreting seismic data. ASKO samples are also different from Boliden samples being hand specimen and stored in surface conditions longer than the other samples. This may have had an impact on their mechanical behaviour and thus P-wave velocities measured in the laboratory.

Reflection coefficients between different rock types

MS	SMS	SP	SKA	TRESKA	QTZR	CRBR	SS	CHLS	BS	SULBS	MCAS	TRECS	CRB
MS	0,00	0,12	0,09	0,09	0,12	0,08	0,19	0,10	0,20	0,16	0,20	0,04	0,06
SMS		0,12	0,09	0,09	0,11	0,08	0,18	0,10	0,19	0,15	0,19	0,04	0,06
SP			0,03	0,03	0,01	0,04	0,07	0,02	0,08	0,04	0,08	0,08	0,06
SKA				0,00	0,02	0,01	0,09	0,01	0,10	0,06	0,11	0,05	0,03
TRESKA					0,02	0,01	0,09	0,01	0,10	0,06	0,11	0,05	0,03
QTZR						0,03	0,07	0,02	0,08	0,04	0,09	0,08	0,05
CRBR							0,11	0,02	0,12	0,08	0,12	0,04	0,02
SS								0,09	0,01	0,03	0,01	0,15	0,12
CHLS									0,10	0,06	0,10	0,06	0,04
BS										0,04	0,00	0,16	0,13
SULBS											0,04	0,12	0,09
MCAS												0,16	0,14
TRECS													0,02
CRB													

Table 9: Reflection coefficients (R) between the Boliden sample set rock types. Coefficients R have been calculated from the seismic impedance values I presented in Table 8 with the formula $R = (I_A - I_B)/(I_A + I_B)$, where $I_A > I_B$. This is the Zoeppritz equation for a normally incident ray (Kearey 2002). In the table, entries with a reflection coefficient equal or greater than 0.06 have been highligted. These are the contacts that can produce a detectable reflection.

5.3 Porosity

The porosities of all samples are presented in Figure 13. In general, the porosities are low. For majority of the COGITO-MIN samples, it remains below 1%. The higher porosity values for ore samples from Luikonlahti (LLAS, LLKU) can be explained by these being hand specimen and stored in surface conditions for a long time, thus their porosity is probably partly due to sampling procedure and weathering of samples in dry, oxidizing and low pressure conditions.

Even low porosities can affect the P-wave velocities measured in laboratory, if they are due to micro cracking related to core retrieval (e.g. Elbra et al. 2011). In Figure 14, porosities of Boliden samples are plotted as a function of depth. The depths are approximated real depths, starting depth of the drill hole plus the location of the sample on the drill core. Dipping of the drill holes has not been taken into account. No tendency of increasing porosities for samples from deeper parts of the drill holes can be seen, thus it seems that either micro cracking due to core retrieval is not significant in these depths, or the effect of it is constant at this depth interval.

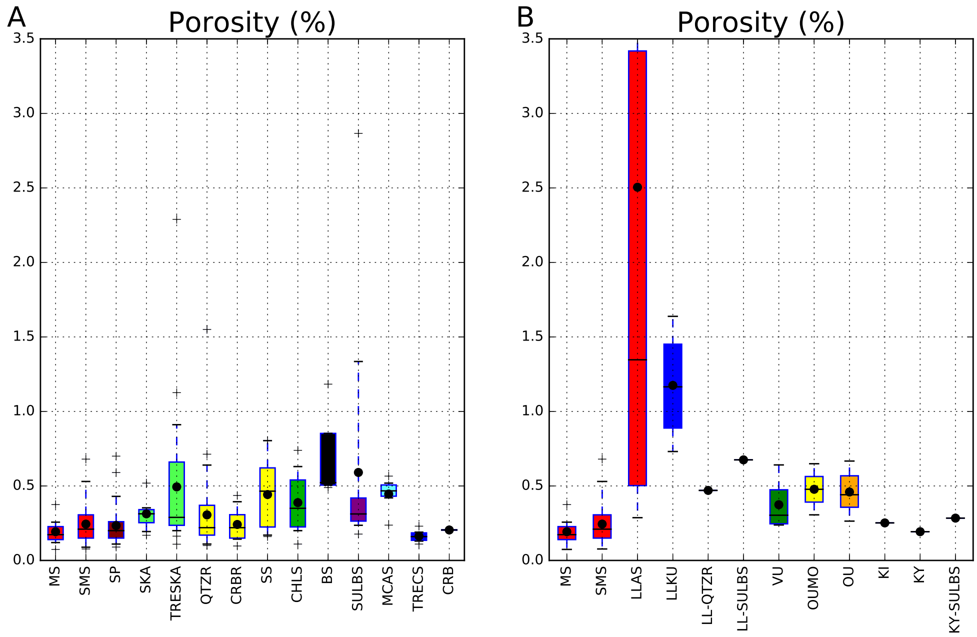

Figure 13: Porosities of samples. Results for Boliden sample set are on the left and for ore sample set on the right. Rock-type abbreviations are listed in tables 4 and 5.

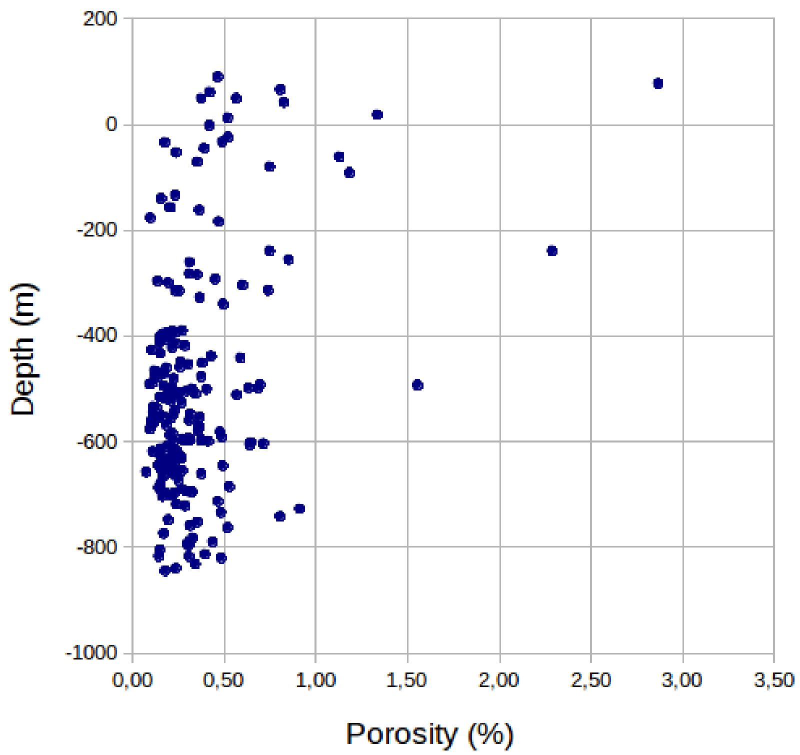

Figure 14: Porosity of Boliden samples as a function of depth.

P-wave velocities as a function of porosity for the the Boliden samples and the ore samples are plotted in Figures 15 and 16. The tendency of decreasing P-wave velocity with increasing porosity can be seen, but variation caused by differences in mineralogical composition seems to be more important than the effect of porosity.

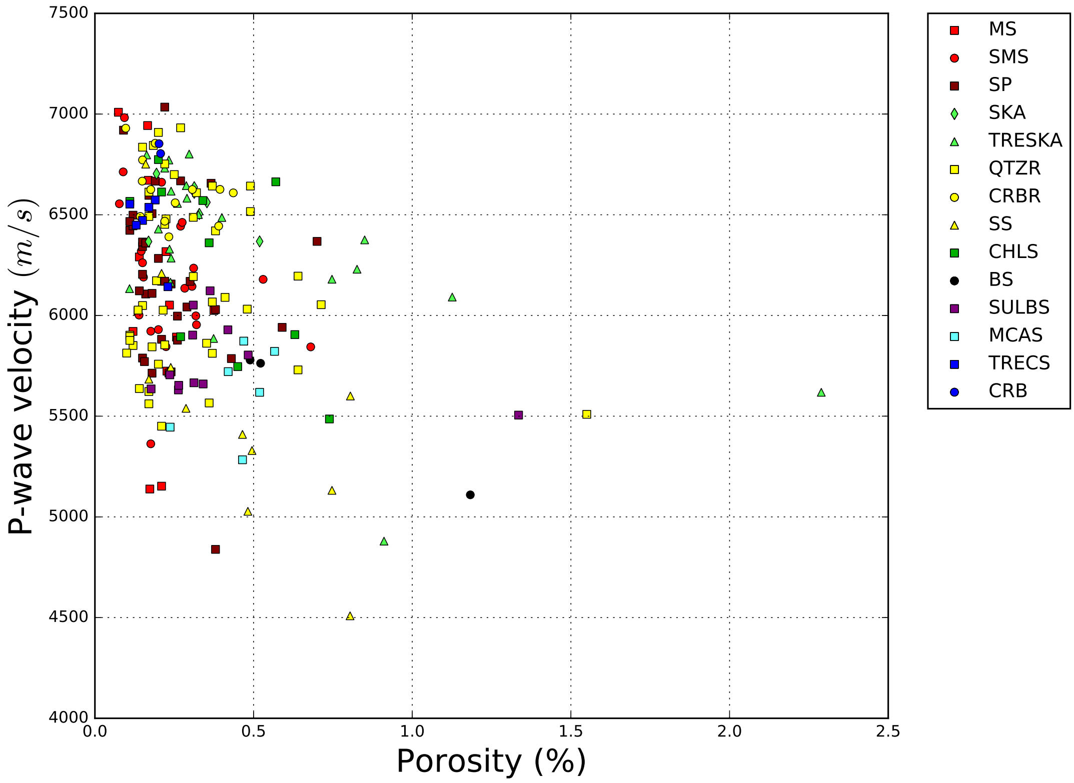

Figure 15: P-wave velocity as a function of porosity for Boliden samples. Rock-type abbreviations are listed in Table 4.

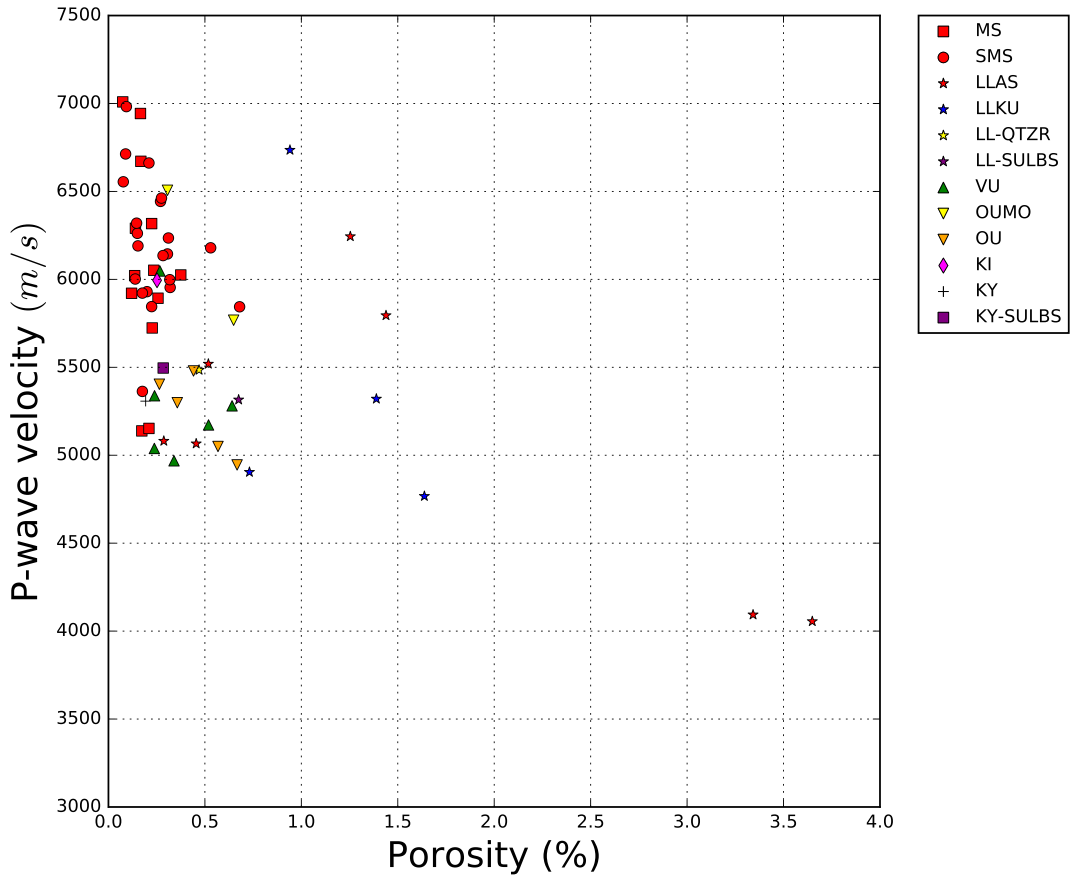

Figure 16: P-wave velocity as a function of porosity for ore samples. Ore-type abbreviations are listed in Table 5

5.4 Magnetic properties

Magnetic susceptibilities and intensities of natural remanent magnetization of samples are shown in Figures 17 and 18, respectively. In the case of magnetic properties, range of values is usually wide and thus the results are represented on a logarithmic scale.

Metamorphosed Precambrian rock samples are usually divided into two populations based on their magnetic susceptibilities, the weakly magnetic and the strongly magnetic (Airo and Säävuori 2013). Susceptibility of the weakly magnetic population is mostly due to paramagnetic mafic silicates and this population is also called paramagnetic population. The strongly magnetic population is often called ferrimagnetic although high susceptibility can also be due to ferro- or antiferromagnetic minerals. The division between these two groups has been set to 2000 µSI in the GTK petrophysical database. This division to two populations can be seen in susceptibility values of COGITO-MIN samples. Therefore average values may be less useful parameters than medians or the colored bar showing the range of values for half of the samples.

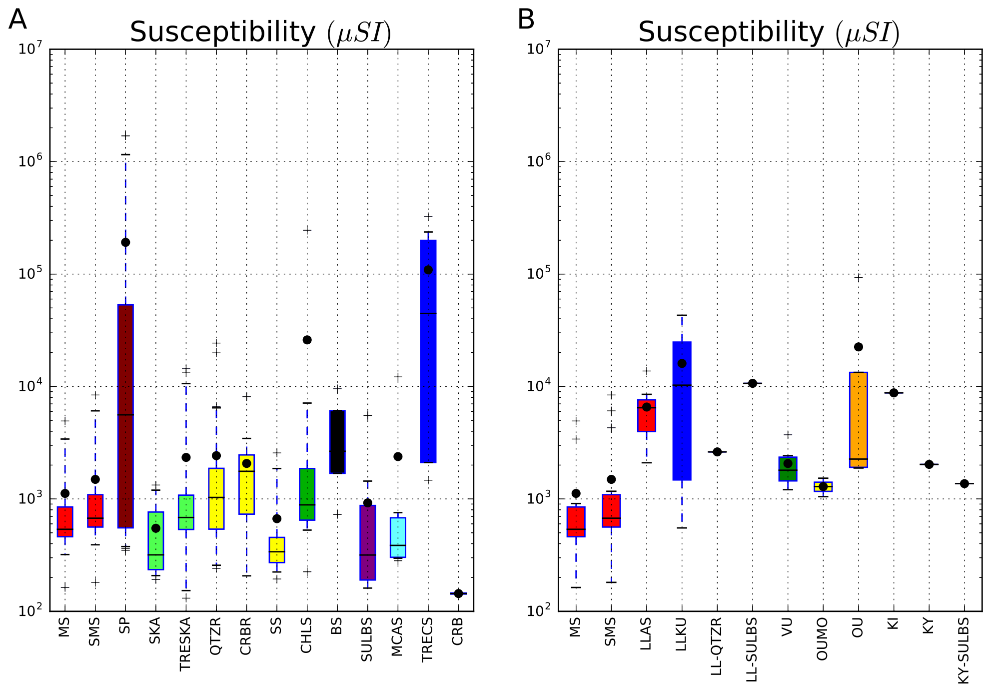

Figure 17: Magnetic susceptibilities of samples, Boliden sample set on the left and ore sample set on the right. Rock-type abbreviations are listed in Table 4 and ore-type abbreviations in table 5.

As can be seen from Figure 17A, most Boliden samples belong to paramagnetic population as their susceptibilities are below 2000 µSI. All rock types except skarn (SKA) have also samples belonging to ferrimagnetic group. These have susceptibilities over 2000 µSI. The only clearly ferrimagnetic rock types are serpentinite (SP) and tremolitic calc-silicate rock (TRECS). Quartz rocks (QTZR) have surprisingly high susceptibility values. Quartz itself is a diamagnetic mineral and has negative susceptibility (e.g.

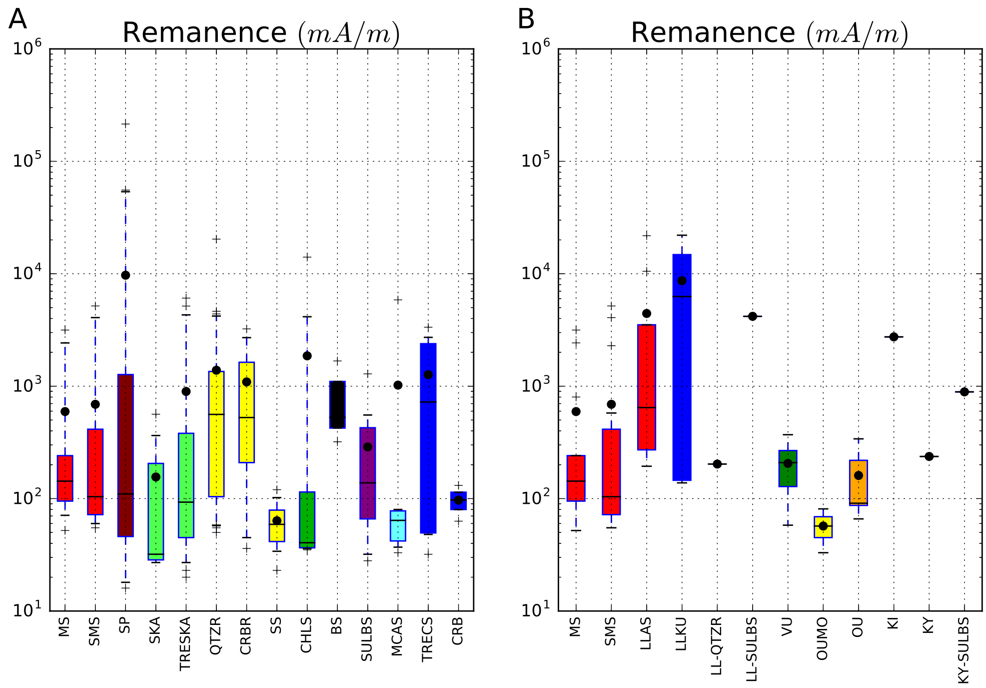

Figure 18: Intensities of natural remanent magnetization of samples, Boliden sample set on the left and ore sample set on the right. Rock-type abbreviations are listed in Table 4 and ore-type abbreviations in table 5.

Schön 2015). High susceptibility values of Kylylahti quartz rocks are due to disseminated sulphides.

Differences in ore mineralogy are reflected in susceptibilities. Pyrite, the dominating ore mineral in Kylylahti, is diamagnetic. Monoclinic pyrrhotite is ferrimagnetic but hexagonal pyrrhotite antiferromagnetic. Magnetite is ferrimagnetic. Rocks rich in pyrite or hexagonal pyrrhotite are excepted to be weakly magnetic and rocks containing monoclinic pyrrhotite or magnetite strongly magnetic. The changing ratios of these ore minerals could thus be evaluated by the susceptibility values. Magnetic behaviour of the ore forming minerals can also be taken into advantage when processing the ore (Pearce et al. 2006).

Kylylahti ore samples (MS, SMS) were from quite limited area. None of the samples was from the deep ore Wombat which contains the most magnetite-rich sulphide mineralization parts in Kylylahti, thus the variability in the susceptibilities of Kylylahti ore samples is quite small. Kylylahti ore samples seem to be from the pyrite-rich part of the ore as is also marked on the logging information of several ore samples. The few Kylylahti ore samples belonging to ferrimagnetic group contain some monoclinic pyrrhotite or, less likely, magnetite. Mineralogical difference between the mining sites in the Outokumpu district are clearly reflected in the susceptibility data (Figure 17B). The samples from sites with pyrite-rich ore, Kylylahti (MS, SMS) and Outokumpu Mökkivaara (OUMO), belong to the paramagnetic population whereas the samples from sites with pyrrhotite-dominated ore, Outokumpu (OU), Luikonlahti (LLAS, LLKU) and Vuonos (VU), show higher susceptibility values.

From Figure 18 it is evident that all rock types in Kylylahti, and also the ore samples from the other sites in the Outokumpu district, carry notable remanence, which should be considered when interpreting magnetic data from the area. The remanence itself is not as important for this work as are the Q ratios (Figure 19) calculated using susceptibility and remanence of the samples.

Q ratios reveal the type of magnetic minerals in the rocks. This parameter is meaningful only for samples in the ferrimagnetic population, thus it was calculated for samples having susceptibility higher than 2000 µSI. Based on petrophysical characterization of Finnish bedrock (Airo and Säävuori 2013) and the results of this work (Figure 19), it can be said that Kylylahti serpentinites (SP) and tremolitic calc-silicate rocks (TRECS) contain coarse-grained magnetite. They have high susceptibility values, but low Q ratios, on average below one.

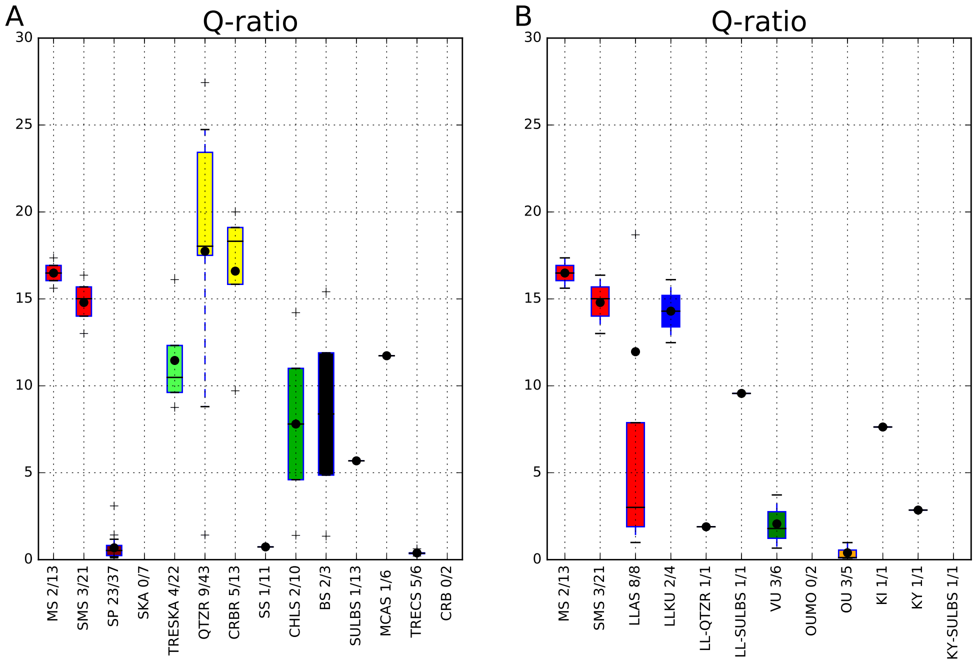

Figure 19: Q ratios of samples having susceptibilities higher than 2000µSI, Boliden sample set on the left and ore sample set on the right. The number of samples with high enough susceptibility and the total number of samples for each rock or ore type are marked after the rock- or ore-type abbreviatios. Rock and ore-type abbreviations are listed in tables 4 and 5.

Q ratios higher than 10 are related to samples containing monoclinic pyrrhotite or fine-grained magnetite (Airo and Säävuori 2013), thus pyrrhotite is dominant sulphide in Kylylahti sulphide disseminations as quartz rock (QTZR) and tremolitic skarn (TRESKA) samples with high susceptibilities (Figure 17A) have also high Q ratio values (Figure 19A). In black schists (BS) and chlorite schists (CHLS) the Q ratios are also comparatively high being between ca. 5-12. These rocks contain some pyrrhotite as was already stated for Kylylahti black schists in the rock type descriptions.

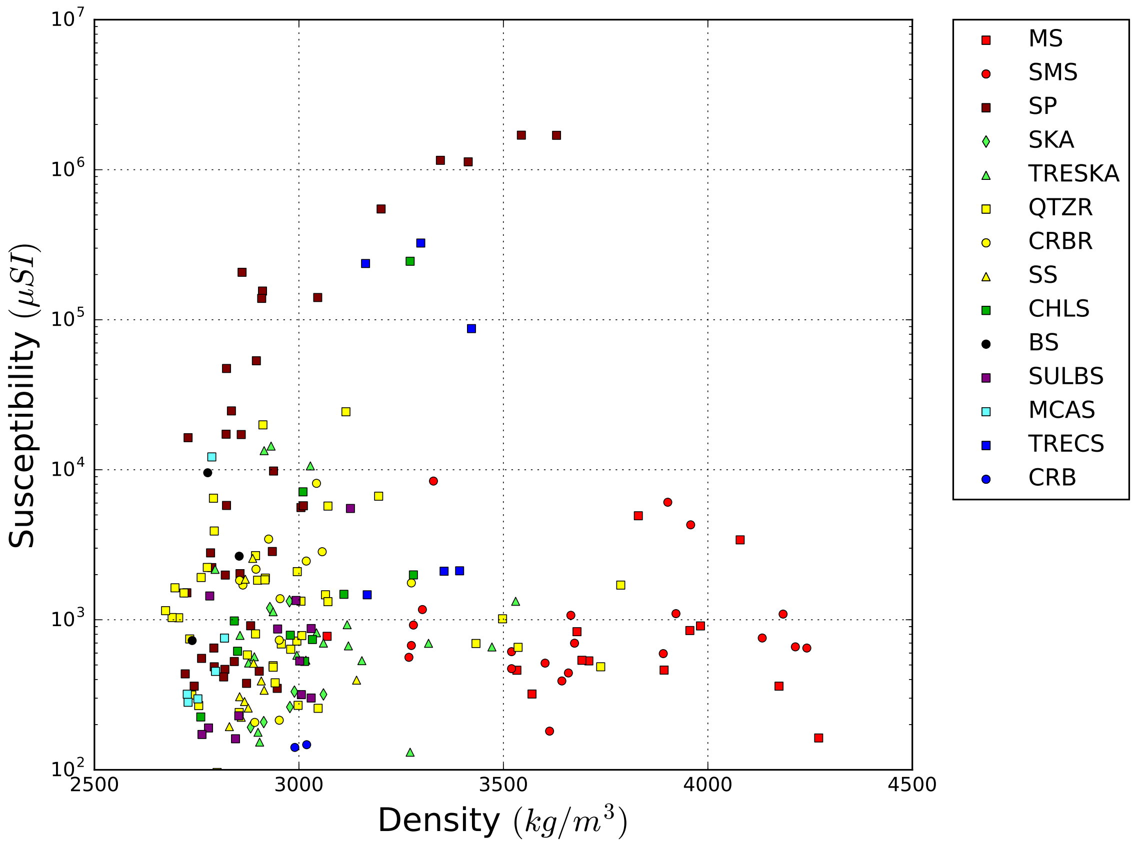

Figure 20: Magnetic susceptibility as a function of density for Boliden samples.

Q ratios of ore samples (Figure 19B) confirm that the ore samples with high susceptibilities in Kylylahti (MS, SMS) and Luikonlahti Kuparivuori (LLKU) are indeed the ones containing pyrrhotite or fine-grained magnetite. With Kylylahti ore samples it has to be noted, that only a few of them belong to ferrimagnetic group. Majority of the Kylylahti ore samples are pyrite rich. Most of the Luikonlahti samples can be attributed to contain pyrrhotite, but based on low Q values, some coarse-grained magnetite is also present. This must also be the case for Outokumpu and Vuonos samples having relatively high susceptibilities but low Q ratios.

Q ratio can also be used when evaluating how dominant the remanence is for the magnetic anomaly related to the rock unit. If Q ratio is bigger than one, remanence dominates and it is important to consider the direction of magnetization in the magnetic data interpretation. The samples used in this work were not oriented and thus the direction of remanence could not be measured. Quite many rock types in Kylylahti have high Q ratios (Figure 19A), thus it would be beneficial to know the dominant direction of remanence when doing magnetic modelling.

When samples are examined in the susceptibility-density plots, as in Figure 20 for Boliden samples and Figure 21 for ore samples, we can see, that only ore samples are clearly separated as their own group, due to their high densities.

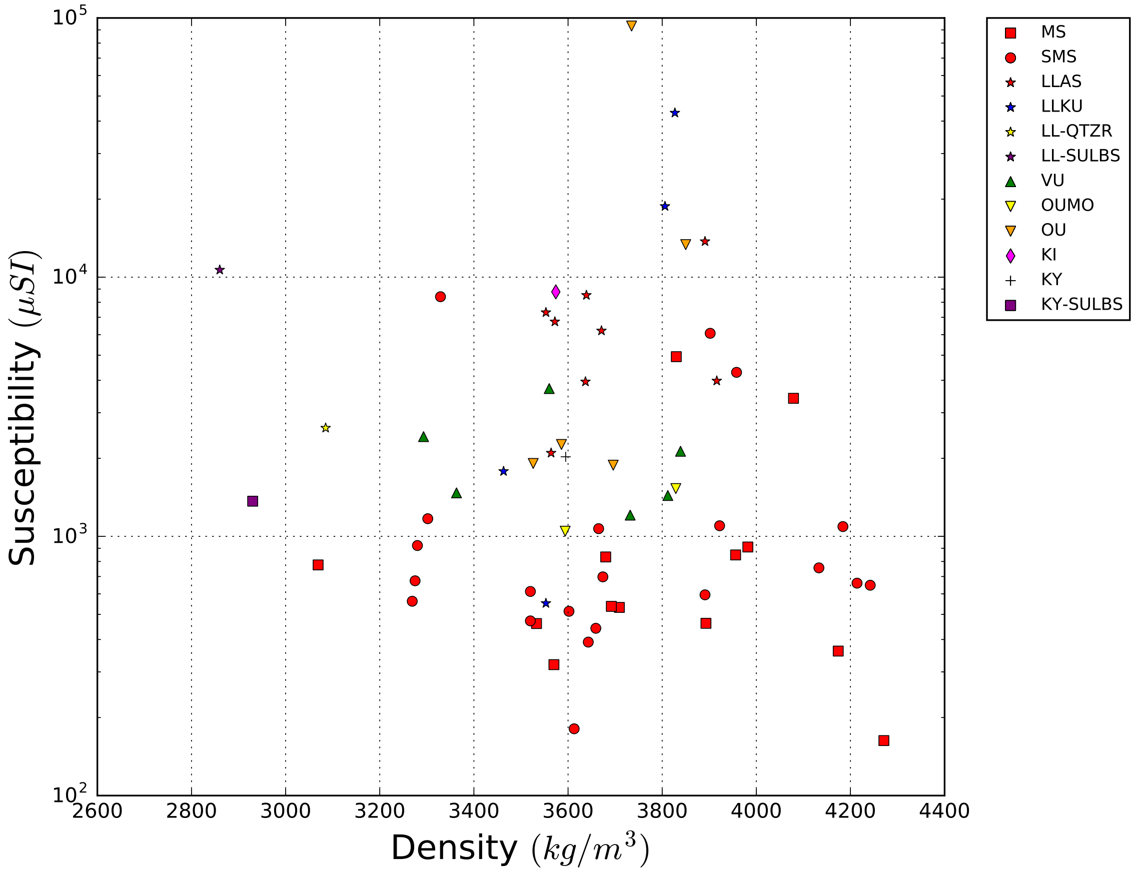

Figure 21: Magnetic susceptibility as a function of density for ore samples.

Figures 22 and 23 show susceptibilities of ferrimagnetic samples (susceptibilities higher that 2000µSI) as a function of Q ratios. Samples containing coarse-grained magnetite can be seen as their own group, having Q ratios around 1. Samples containing monoclinic pyrrhotite form their own group with Q ratios higher than about 10 (Figure 22).

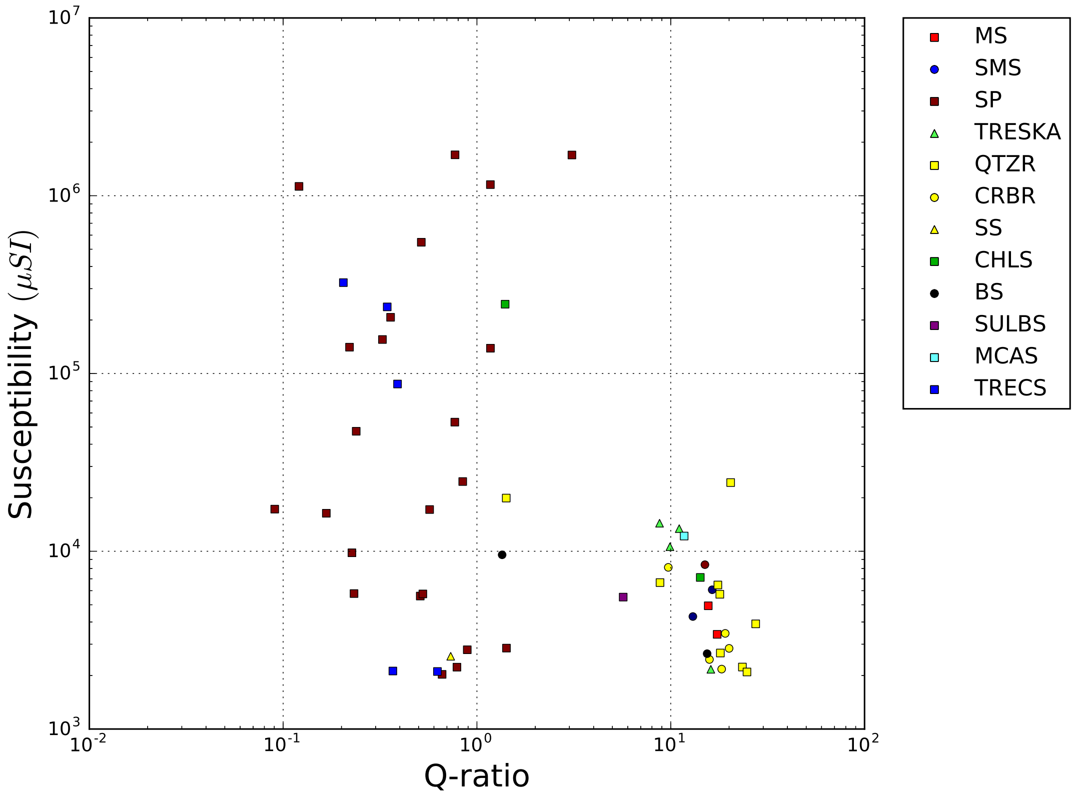

Figure 22: Magnetic susceptibility as a function of Q ratio for Boliden samples. Only samples with susceptibilities higher than 2000µSI are shown.

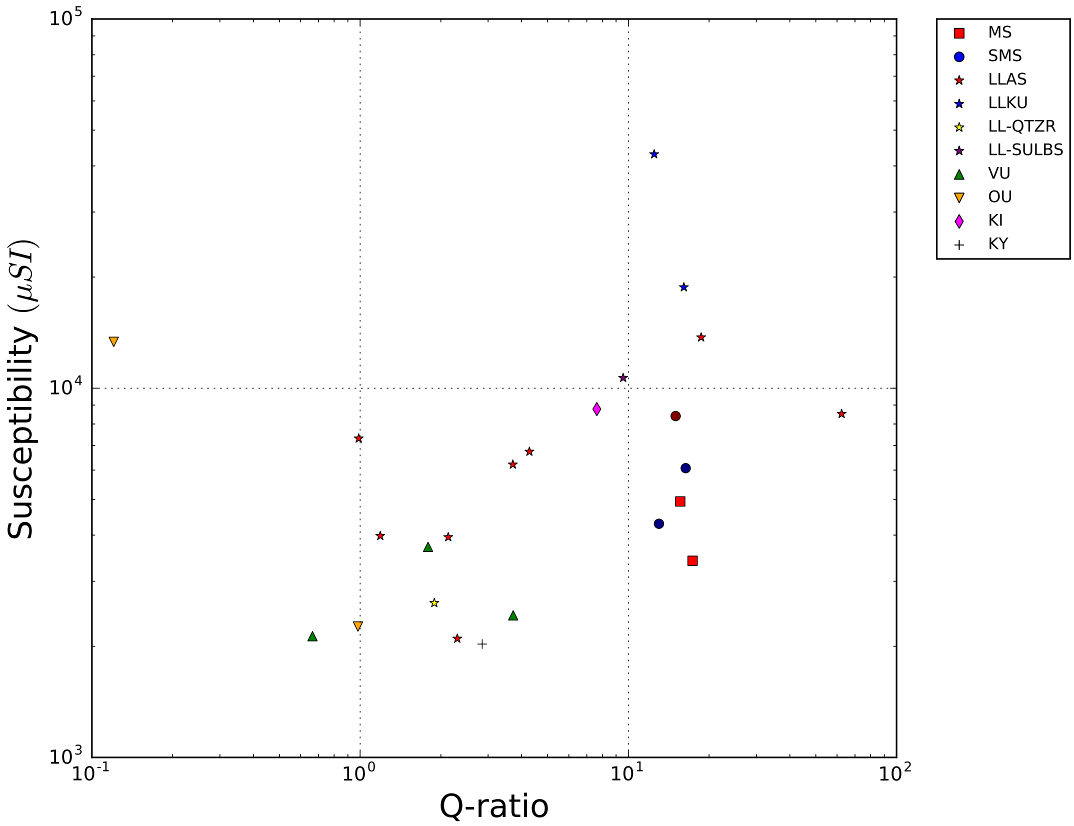

Figure 23: Magnetic susceptibility as a function of Q ratio for ore samples. Only samples with susceptibilities higher than 2000µSI are shown.

5.5 Electrical properties

In Figures 24 and 25 are specific resistivities and conductivities measured by inductive method. Inductive method was developed for measuring resistivity of highly conducting samples, thus most common rocks do not get any resistivity value by inductive method. If a sample was not measurable by inductive method, it's conductivity has been set to zero to get more representative average values for conductivity of rock types. Conductivity is the reciprocal of resistivity. It is still useful to see the conductivity values separately. In those, all samples can be included, thus the rock type averages are more meaningful. It also shows, which rock types are likely to produce a signal, when searching conductive rocks by electric or electromagnetic survey methods.

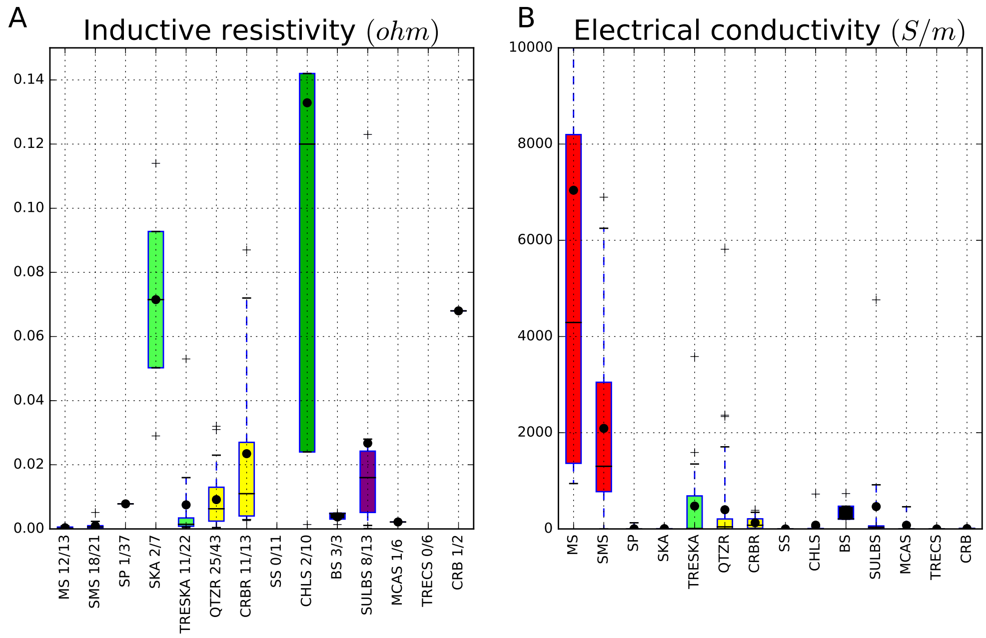

Figure 24: Inductive resistivities and conductivities of Boliden samples.Number of samples measurable by inductive method and total number of samples are marked for each rock type. The electrical conductivity plot contains all samples. Rock-type abbreviations are listed in Table 4.

Most of the Outokumpu assemblage rocks, serpentinites (SP), soap stones (SS), skarns (SKA, TRESKA), chlorite schists (CHLS) and tremolitic calc-silicate rock (TRECS) are practically non-conductive (Figure 24A) and gave only a few results when measured by inductive method. Those were also the highest resistivities measured by that method. Naturally ore samples (MS, SMS) are conductive 24B), but also quartz (QTZR) and carbonate rocks (CRBR, CRB) have many conductive samples due to their disseminated sulphide content.

According to their magnetic properties, serpentinites (SP) and tremolitic calc-silicate rocks (TRECS) contain substantial amounts of magnetite. Magnetite itself is a conductor, but when it is found in coarse grains, as in Kylylahti, the grains are usually not connected well enough for the rock to become

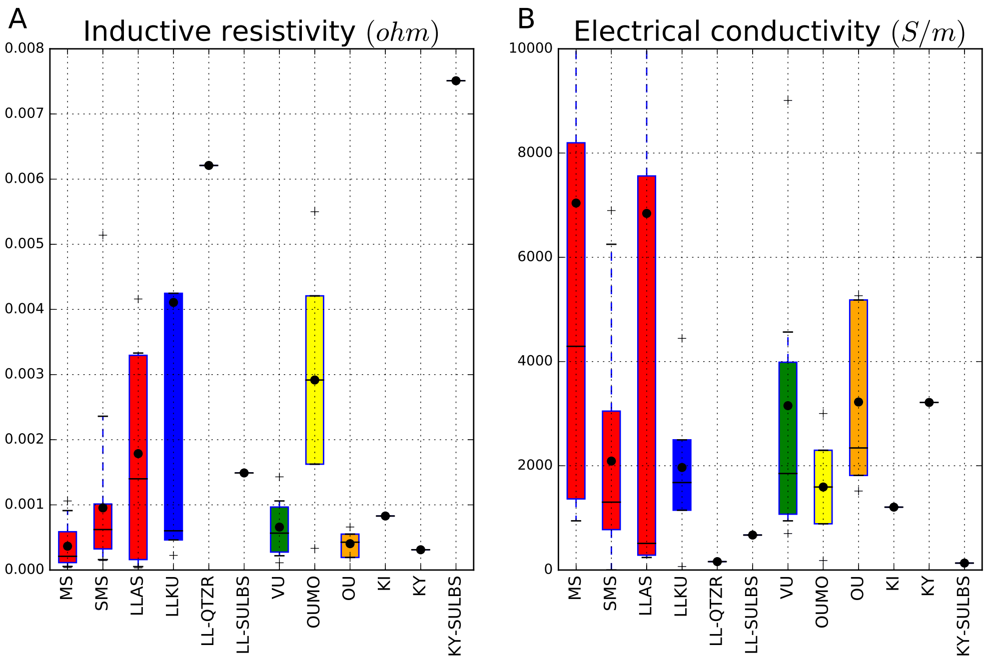

Figure 25: Inductive resistivities and conductivities of ore samples. Ore-type abbreviations are listed in Table 5.

conductive. Black schist (BS) and sulphide-rich black schist (SULBS) contain sulphides but also graphite that makes them conductive (Figure 24B).

As would be expected, almost all ore samples were measurable with inductive method and thus conductive (Figure 25). Only four samples from Kylylahti and one sample from Luikonlahti Asuntotalo got no reading. Massive ore is the most conductive rock type, also an expected result.

As inductive method is the most suitable for conductive rocks, rocks with higher resistivities are better measured by galvanic method. Specific resistivities of Boliden samples, measured by galvanic method, and IP estimates, calculated with the equation 4.11.4, can be seen in Figure 26. For ASKO sample set no galvanic specific resistivity values are reported, because those samples were too conductive to be measured with this method.

Almost all Boliden samples were measurable with galvanic method. In total 15 samples, mainly ore (MS, SMS) or black schists (BS) were too conductive to be measured with it. Figure 27A shows that the ore samples have the smallest resistivities. The relatively high resistivity and high chargeability, seen in the high values of IP estimates (Figure 27B), reveals that the sulphide mineralizations in these samples form more disseminated patches than continuous veins. Most other Outokumpu assemblage rocks have resistivities in order of typical silicate bearing rocks, but some samples have disseminated sulphides diminishing the resistivity. Due to highly conductive graphite, the resistivities of black schists (BS) and sulphide-rich black schists(SULBS) are low. Sulphide disseminations in sulphide-rich black schists cause

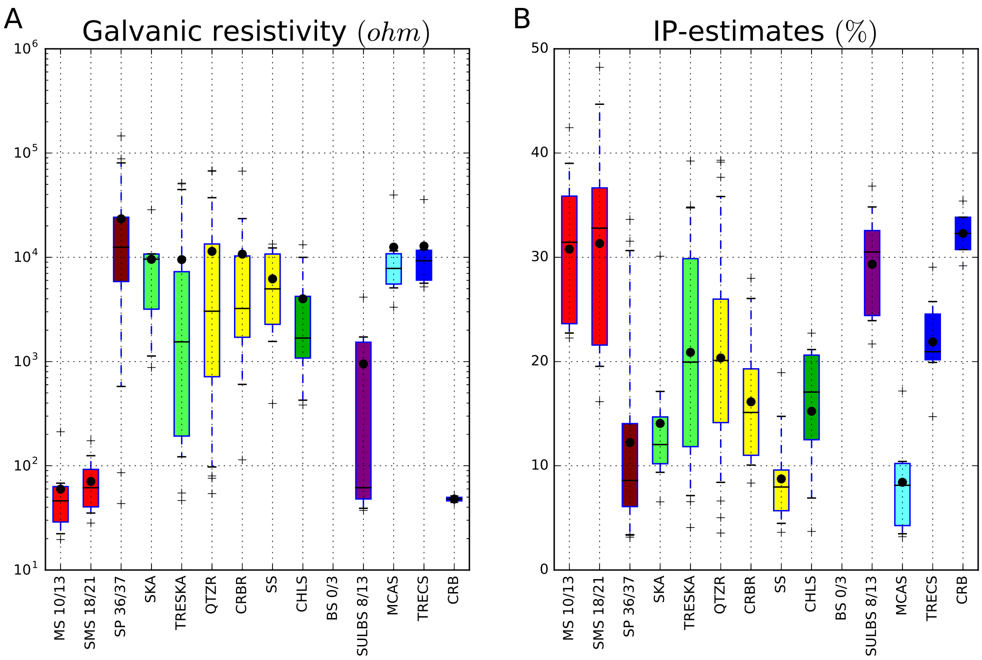

Figure 26: Galvanic resistivity values and IP estimates for Boliden samples. Rock-type abbreviations are listed in Table 4. After the rock-type abbreviation, number of samples measurable by galvanic method and total number of samples are marked for each rock type. If no number is marked, all samples of that rock type were measurable.

its high IP values as is also the case for other rocks containing sulphide disseminations, especially some samples of carbonate-skarn-quartz rocks (CRBR, TRESKA, QTZR). In serpentinites (SP) and tremolitic calc-silicate rocks (TRECS) the chargeability is due to their magnetite content.

Galvanic resistivities of Boliden samples are plotted as a function of IP estimates in Figure 27. The chargeability of ore samples (MS, SMS) can be clearly seen, as well as a large number of other samples containing disseminated sulphides and thus having high chargeability.

Galvanic resitivity and porosity of Boliden samples are plotted on Figure 28. Porosity can decrease the resistivity of rocks rapidly, due to pore fluids that can be charge carriers. In the Kylylahti rocks porosities are so low, that the differences in resitivities are mainly due to differences in mineralogy.

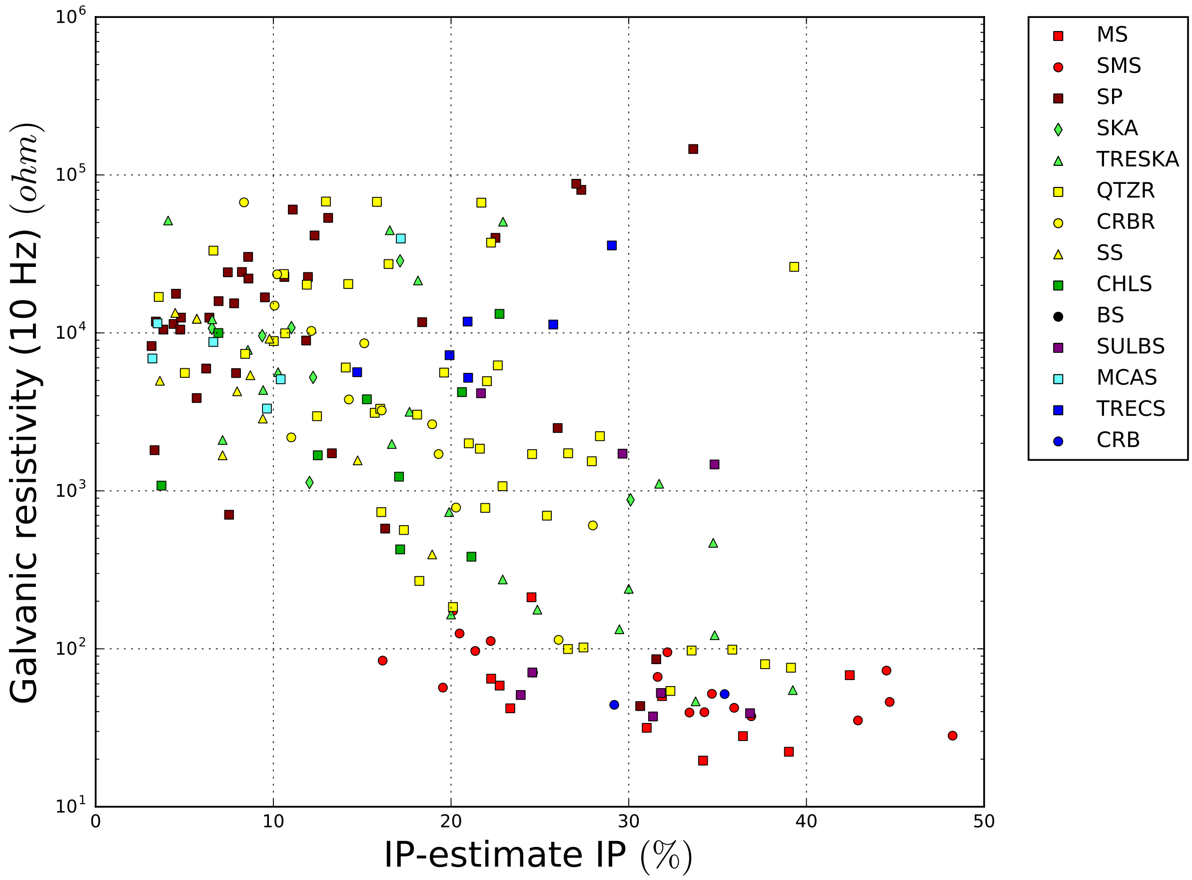

Figure 27: IP estimate IP vs Galvanic resistivity for Boliden samples that were measurable by galvanic method. Rock-type abbreviations are listed in Table 4.

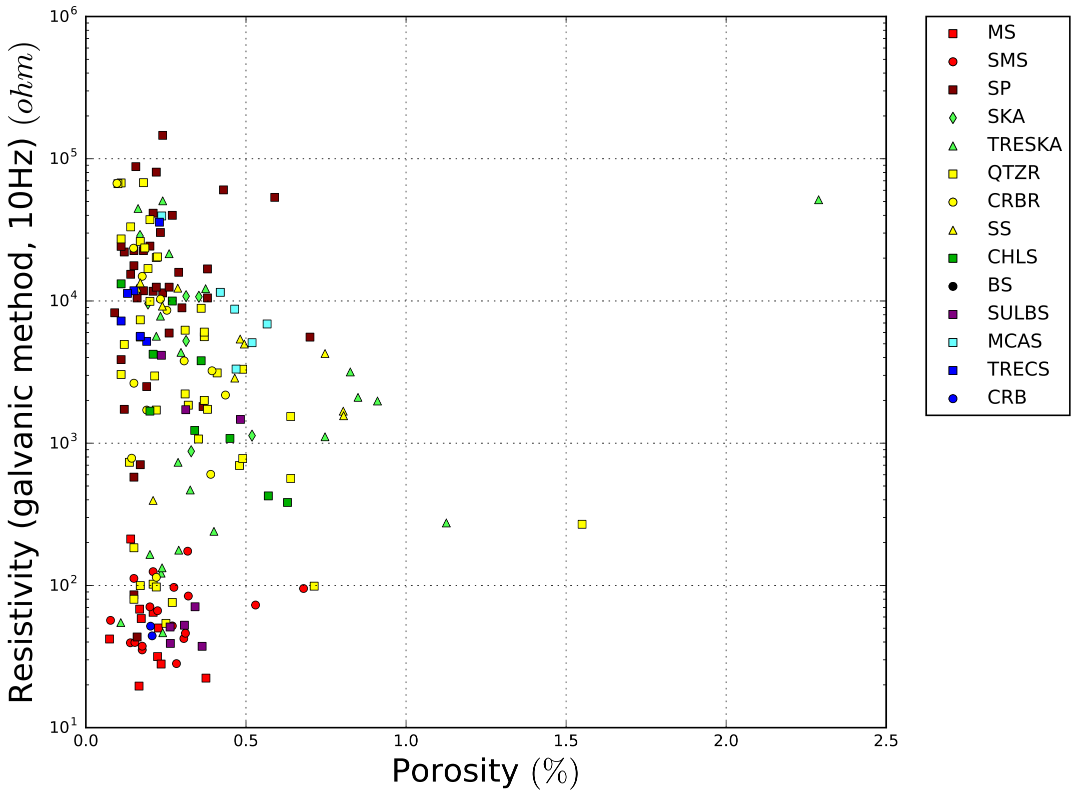

Figure 28: Specific resistivity as a function of porosity for Boliden samples. Rock-type abbreviations are listed in Table 4

5.6 Petrophysical properties by drill holes

Responses of geophysical methods are dependent on physical properties of rocks. In addition of knowing the properties of different rock types, it is also good to have an idea of what kind of contacts between the rock types can be expected. Drill core logging data is useful for this purpose. Therefore, in this chapter the petrophysical data previously shown by rock type are shown for each drill core separately.

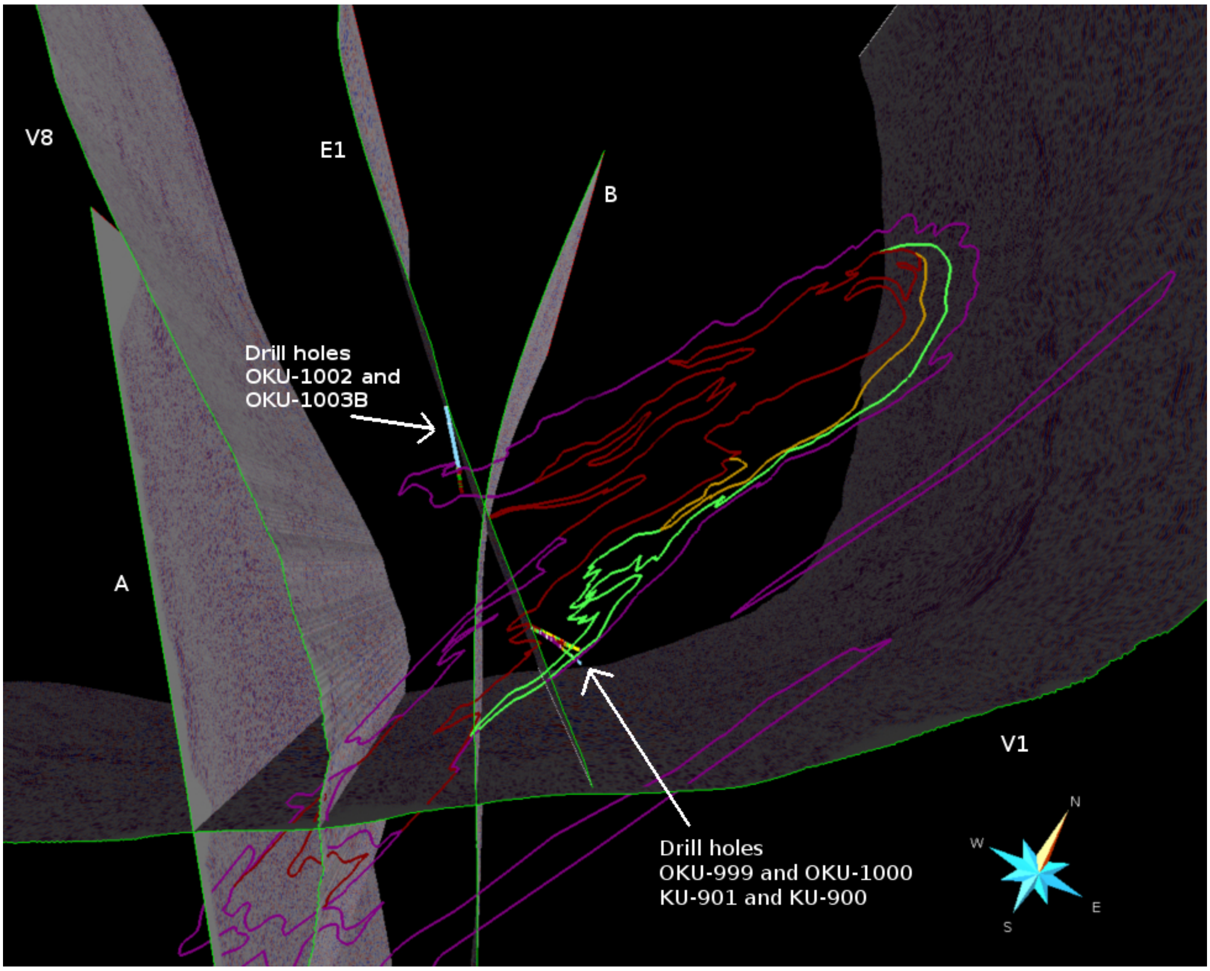

Figure 29: On the left: Seismic lines intersecting Kylylahti massif on a horizontal map view and the location of COGITO-MIN petrophysics drill holes relative to them. Lines V1, V8 and E1 are HIRE lines (Kukkonen et al. 2012a) and lines A and B COGITO-MIN 2D reflection seismic lines (Heinonen et al. 2018). Shape of the Kylylahti massif from the geological map is also shown.

In this work, we have data from six drill cores from Kylylahti. The locations of the drill holes can be seen in Figures 29 and 30. OKU drill holes start from the surface. OKU-999 and OKU-1000 are in the eastern side of the Kylylahti massif, right above the sulphide mineralizations. OKU-1002 and OKU-1003B are situated more to the west, where the massif does not reach the surface. OKU-1002 and OKU-1003B punctuate the Kylylahti massif after a few hundreds of meters of mica schists.

The starting points of drill holes KU-900 and KU-901 are in a mine tunnel, almost 500 metres below the surface (Figure 30). Their starting points are close to each other, but the holes are directed differently. Both go through sulphide mineralizations. The drill hole KU-901 was the principal petrophysical drill hole for measurements in the COGITO-MIN project. It goes through most of the rock types of the Kylylahti massif. Other drill cores provide supplementary samples in order to have a more representative sample set and to have all the relevant rock types covered.

KU-901 is also quite close to the COGITO-MIN vertical seismic profiling (VSP) measurement sites. The COGITO-MIN project involved acquisition and integration of multi-scale seismic methods including surface-based 2D and 3D active-source and passive seismic measurements and an underground in-mine VSP survey using conventional and fiber-optic technologies (Riedel et al. 2018). One VSP seismic section will be presented later in this work, when discussing application of measured seismic impedances in interpretation of seismic data from Kylylahti (Chapter 6.5). In Figure 29 are also shown several seismic lines. Lines V1, V8 and E1 are HIRE lines (Kukkonen et al. 2012a). Line E1 will be used when discussing the results in Chapter 6.5. Lines A and B are COGITO-MIN 2D reflection seismic lines (Heinonen et al. 2018). The results of this work were used when processing COGITO-MIN seismic data.

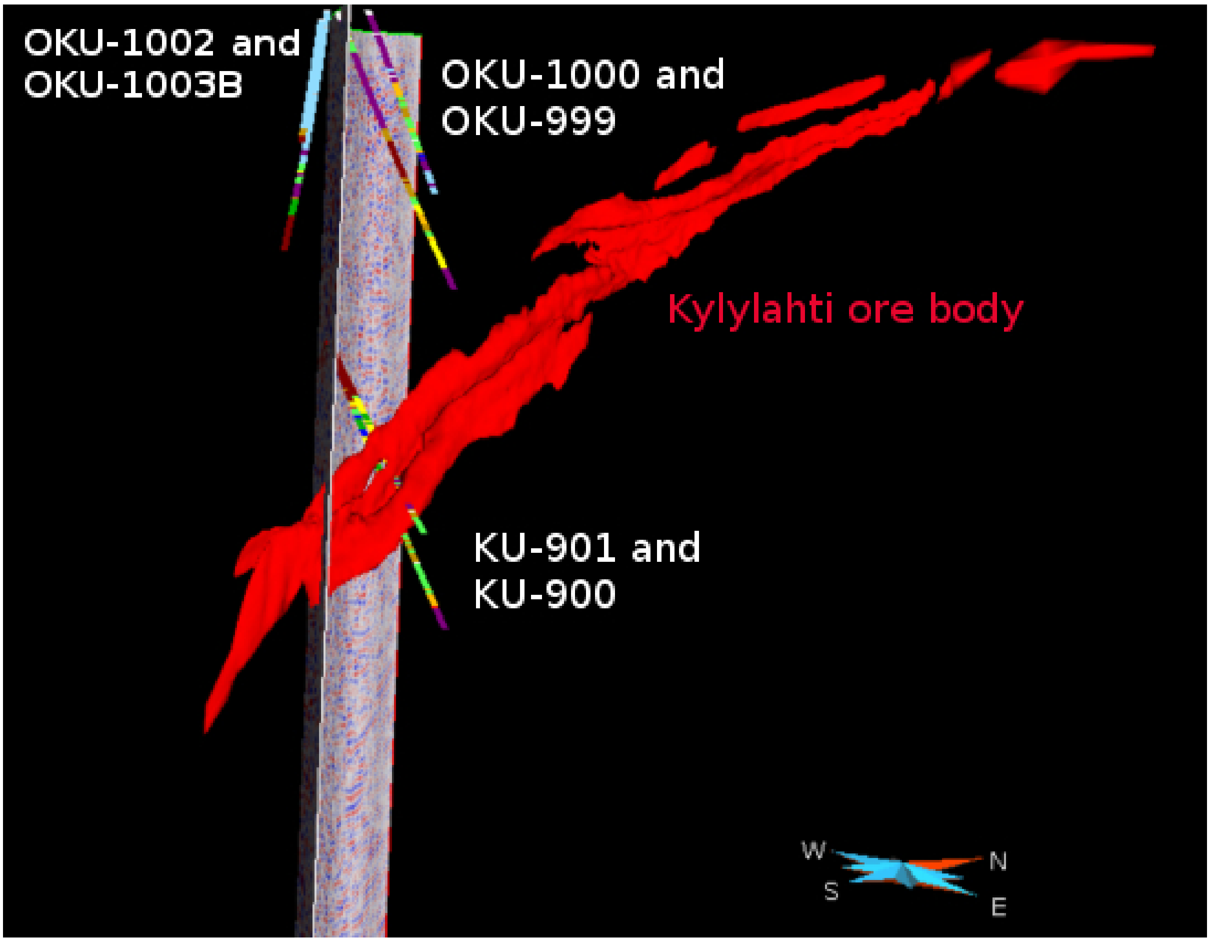

Figure 30: Seismic line E1 (Kukkonen et al. 2012a), COGITO-MIN petrophysics drill holes and the Kylylahti ore body, seen from the side towards northwest.

Figures 31 and 32 are legends for rock types and lithologies, respectively, used when plotting the drill core data. It is to be noticed that not all of them are represented in the sample sets of this work, listed in Table 4. Rock types and lithologies of the drill cores have been determined by Boliden geologists and can be found from the Kylylahti drill hole database (Boliden 2016b) . Both rock types and lithologies are plotted. Rock type is more detailed classification. Lithologies combine together rock types belonging to the same lithological unit, thus it often clarifies the interpretation. One big advantage is, that disseminated ore has been classified as its own lithology, whereas in rock types disseminated ore samples are classified according to their host rocks. Petrophysical properties of rocks with sulphide disseminations differ from those of their host rocks without dissemination, thus in the interpretation, it is valuable to know, where these disseminations are located.

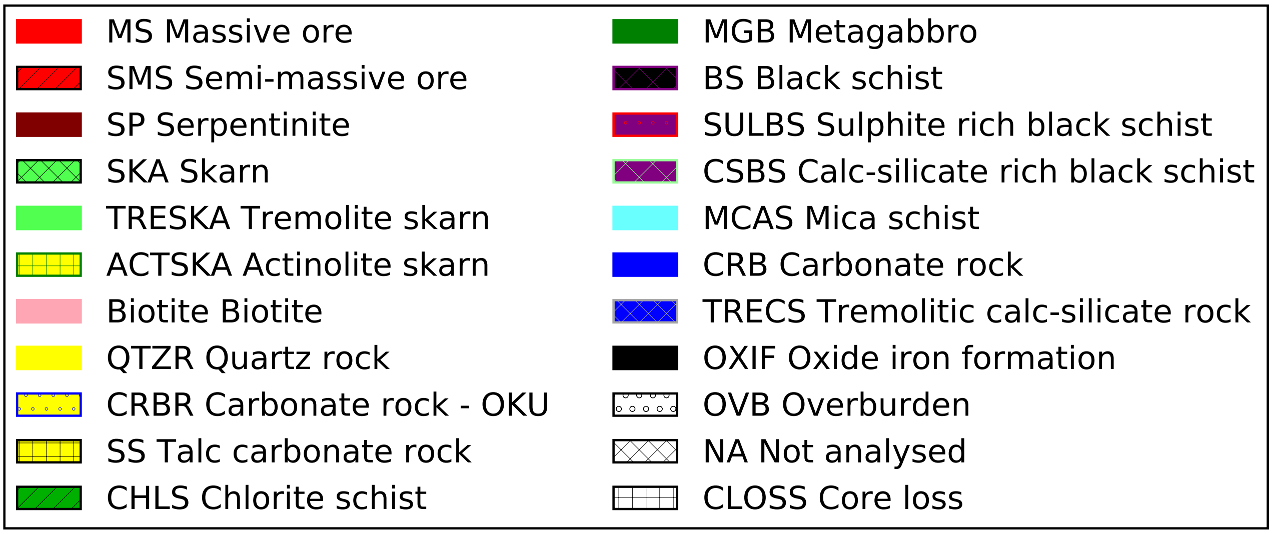

Figure 31: Rock types present in drill holes used in this study. Not all rock types are represented in sample sets. Rock types of drill core samples are listed in Table 4.

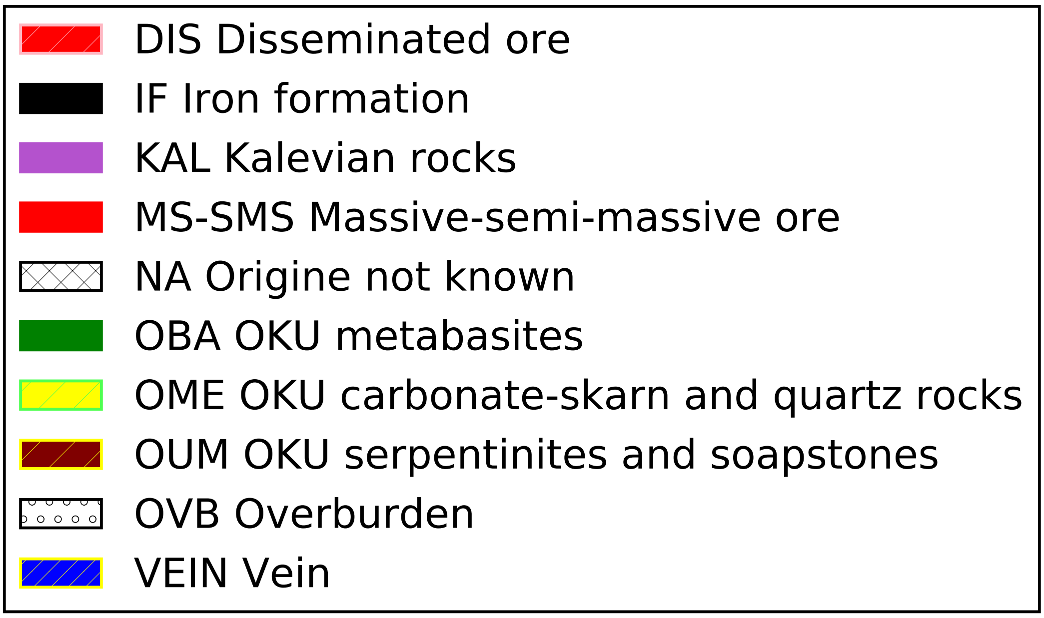

Figure 32: Lithological units present in drill holes used in this study.

Figures 33 to 37 show selected petrophysical parameters for the drill cores; density, P-wave velocity, porosity, conductivity (as a reciprocal of inductive resistivity) and magnetic susceptibility. Depths are relative to the beginning of the drill hole. Starting heights, height coordinate of the starting point, of the drillholes are also marked in the figures. Only the sampled parts of the drill cores are shown.

Figure 33 shows the data for the drill hole KU-901. The drill hole starts from -389.6 m, from the Kylylahti mine tunnel, almost 500 m underground, and is 459.1 m deep. More than a half of the COGITO-MIN samples are from this drill hole. The starting point of the drill hole is inside of the Kylylahti massif and the first hundred metres of it are Outokumpu assemblage serpentinites (SP) with small intervening oxide iron formation (OXIF) starting at 6 metres depth and a carbonate rock (CRB) vein at 78 metres depth. The samples from the iron oxide formation have been classified as serpentinites, by Boliden geologists, which has an impact on the interpretation. This will be discussed in detail later (Chapter 6.2.1).

In Figure 33, at relative depth of 100 metres, close to 600 m underground, starts a typical, complex sequence of Outokumpu assemblage carbonate-skarn-quartz rocks (CRBR, SKA, TRESKA, QTZR). Starting at approximately 120 m, it is interrupted by an approximately 20 m thick layer of Outokumpu metabasites (OBA), which is composed of chlorite schists (CHLS) and tremolite schists (TRECS). After the metabasites, carbonate-skarn-quartz rocks dominate down to about 420 metres depth. There the hole reaches the eastern margin of the Kylylahti massif, and ends in Kalevian rocks (KAL), represented by sulphide-rich black schist (SULBS). Carbonate-skarn-quartz rocks contain disseminated sulphides on several sequences. The drill hole KU-901 goes through the Kylylahti ore (MS, SMS) at around 300 m depth.

The density values (Figure 33) are the highest for massive and semi-massive ore (MS, SMS). Also most of the samples representing disseminated ore have elevated density, as well as the oxide iron formation. The P-wave velocity values are quite scattered. The only clearly contrasting sequence is the low velocity area of soap stones at around 350 metres depth. The P-wave velocities drop clearly when going from Outokumpu assemblage rocks to Kalevian rocks, almost at the bottom of the hole. Seismic impedance values indicate that the contacts between the oxide iron formation and serpentinite (OXIF-SP, 6 m and 17 m depth), serpentinite and carbonate-skarn-rocks (SP-QTZR, 103 m), Outokumpu metabasites and carbonate-skarn-quartz rocks (OBA-OME, 157 m), soap stones and skarns (TRESKA-SS and SS-SKA, between 315 m and 350 m) and the margin of the massif (CRBR-SULBS, 426 m) are possible reflectors.

The ore itself has high seismic impedance, but especially from the top, it is surrounded by sulphide disseminations and clear sharp boundaries are not seen in the data (Figure 33). According to the laboratory results (Chapter 5.2), the ore (MS, SMS) should be detectable against almost any background. On the other hand, several samples have seismic impedances comparable to those of the ore samples

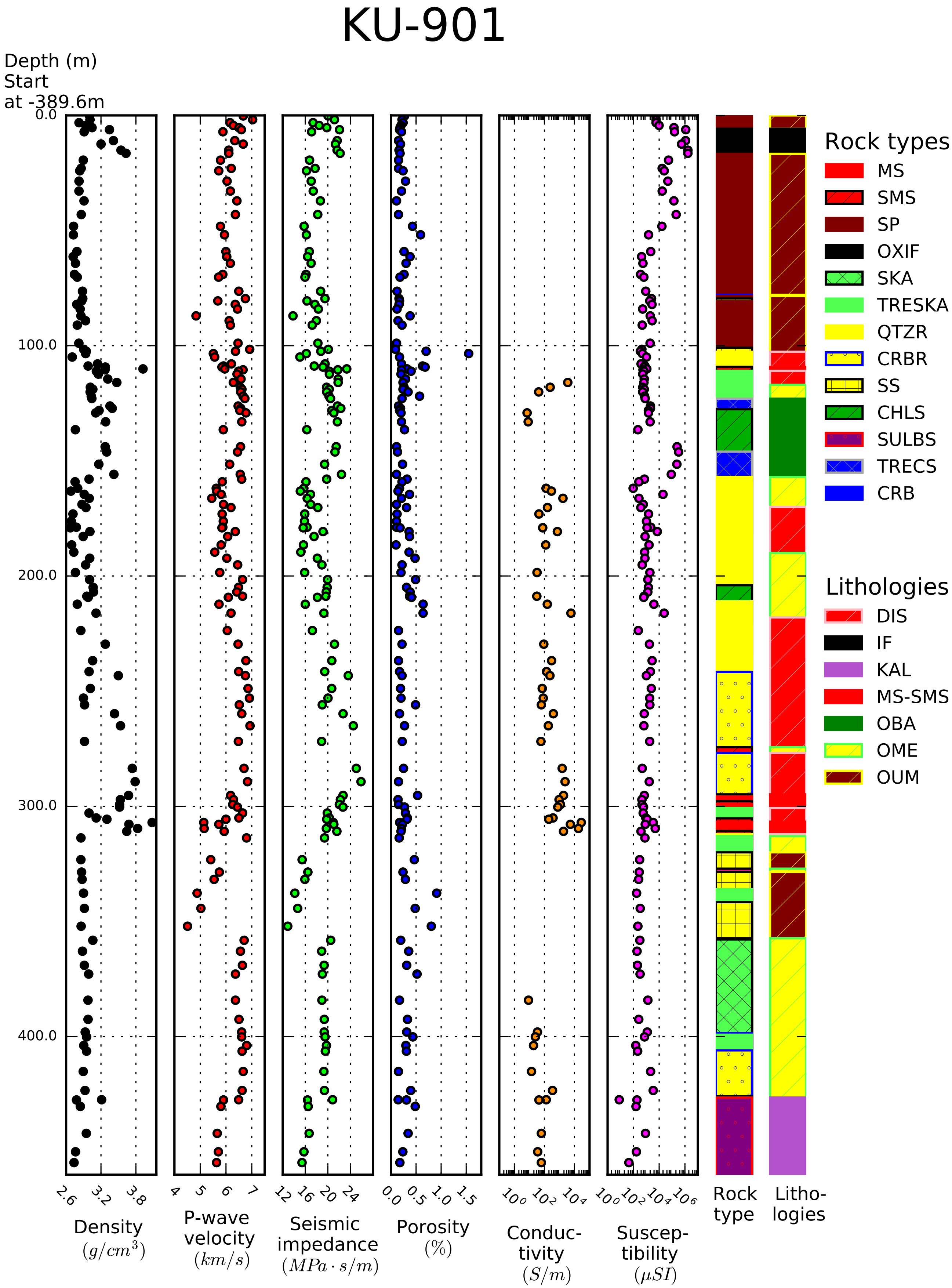

Figure 33: Data for the drill hole KU-901.

(Figure 10). When the rocks in contact with the ore are of this type, there will be no detectable reflection. More dense sampling along the drill core would be needed to resolve the reflectors more thoroughly.

Porosities are very low throughout the hole. The few samples with slightly higher porosities, around 100 and 350 metres are close to areas of brecciated rocks, but do not belong to those. The sample description does not give any more hints for the reason of higher porosity. Elevated porosity could also be due to microcracking during the core retrieval, or other sample handling.

Conductivity values reveal the areas with massive and semi-massive ore (MS, SMS) or disseminated ore (DIS), whereas serpentinites (SP), chlorite schists (CHLS) and soap stones (SS) are not conductive at all. The carbonate rocks (CRBR) and sulphide-rich black schists (SULBS), at the bottom of the hole, are both expected to contain disseminated sulphides, sulphide-rich black schist also graphite, and are thus conductive.

Susceptibilities are high for serpentinites (SP) and tremolitic calc-silicate rocks (TRECS) that both contain magnetite. Although the serpentinites have samples with high susceptibilities, part of the serpentinite samples are clearly paramagnetic. According to Airo and Loukola-Ruskeeniemi (2004), the susceptibility of ultramafic rocks decreases with increasing degree of alteration as biotization and talc-carbonate alteration leads to loss of magnetite. The degree of metamorphism gets higher with depth in Kylylahti, which is illustrated also in the susceptibility values of the serpentinites from KU-901 (Figure 33). The lowest susceptibilities are associated with sulphide-rich black schists (SULBS). This indicates that the main sulphide in these black schists is most probably pyrite.

The drill hole KU-900 (Figure 34) starts also from the mine tunnel. The starting depth is -389.7 m and the hole is 330 m deep. We do not have samples from its lowermost 120 m, nor from the first 200 metres. KU-900 starts practically from the same point as KU-901 (Figure 30) but is directed a bit more upward and goes through the ore from a different point. It provides complementary samples to those from KU-901, most importantly 15 samples of massive or semi-massive ore (MS,SMS), 24 samples in total. According to logging information (Boliden 2016b), all the samples from KU-900 contain sulphides.

The sampled part (Figure 34) of the drill core consists mostly of Outokumpu carbonate-skarn-rocks (OME) and ore (MS-SMS), with two sequences of Kalevian black schists (SULBS, BS). The upper part contains semi-massive ore between 200 and 210 metres, although that part has been logged as quartz rock (QTZR). The drill core continues as quartz rocks (QTZR) and tremolitic skarns (TRESKA) before two black schist intercalations at 237 m and 247 m. The second black schist sequence is in contact with carbonate rocks. The Kylylahti ore body is encountered at about 253 metres depth. The zone with massive and semimassive sulphide mineralizations continues down to 276 m, with two intervening

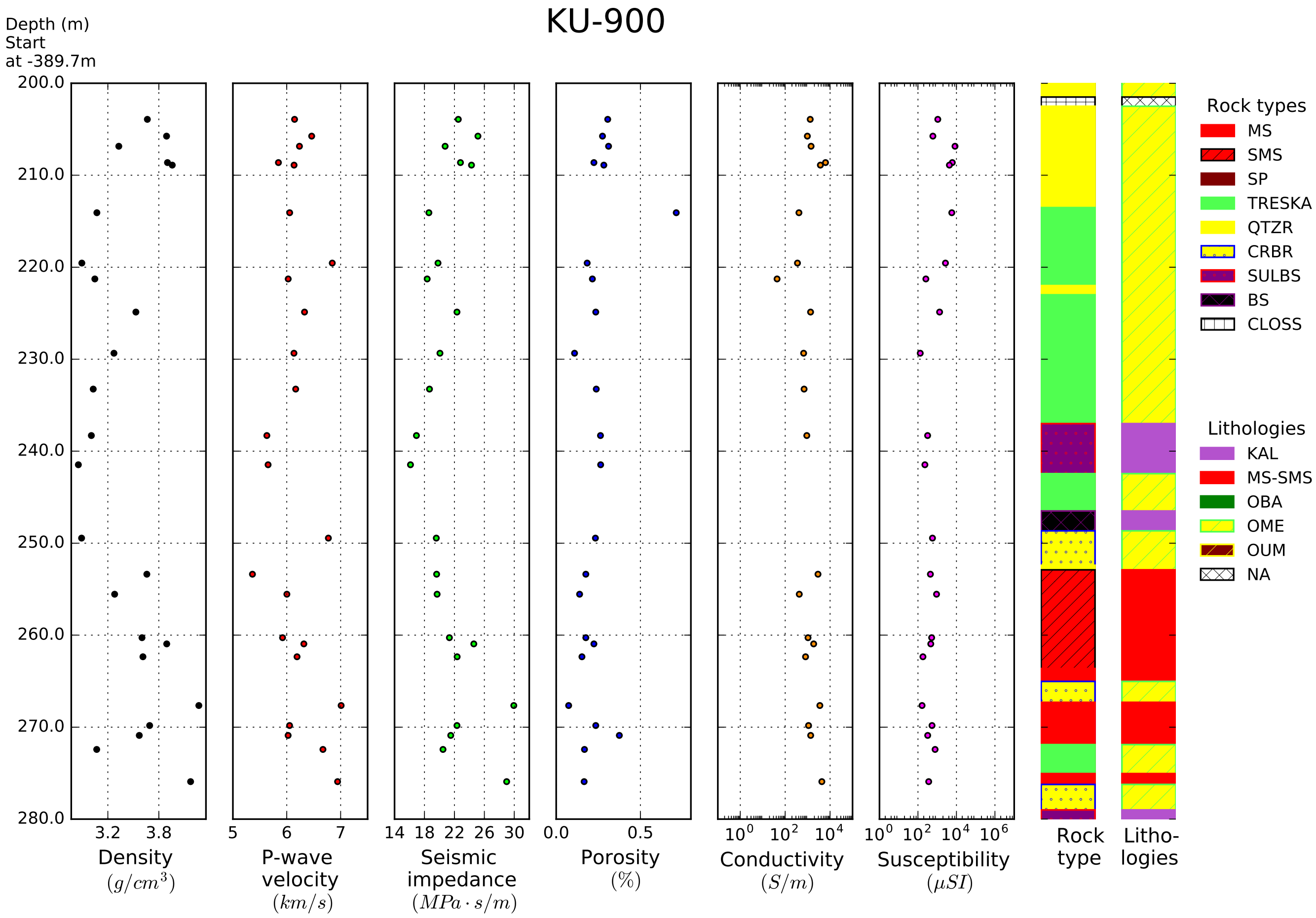

Figure 34: Data for the drill hole KU-900.

sections of carbonate rocks (CRMR, 265 m) and tremolitic skarn (TRESKA, 276 m). In this work, we had no samples from the drill core KU-900 below the ore body.

The density values (Figure 34) correlate with sulphide content, massive ore (MS) having the highest values. In the P-wave velocities of KU-900 samples, no clear tendencies can be seen. The seismic impedance data from KU-900 indicates that the ore (MS or SMS) could be a reflector in contrast with other Outokumpu assemblage rocks, but the sampling density is too low to make any final conclusions on that. Porosity is very low overall. Conductivity values are elevated throughout the sampled part of the drill core, as all the samples contained some sulphides. Susceptibility values are highest in the uppermost part. This is probably due to disseminated pyrrhotite whilst the ore itself is pyrite dominated.

The drill hole OKU-999 (Figure 35), starts from the surface, at the height 94.3 m. The total length of the hole is 464 m. It is situated above the holes KU-901 and KU-900 (Figure 30). It punctuates the Kylylahti massif from its eastern edge. The first 200 metres of the drill core are Kalevian rocks (KAL), black schists (BS) and sulphide-rich black schists (SULBS) intervened by a sequence of Outokumpu assemblage soap stones (SS) and tremolite skarns (TRESKA) between 23 m and 68 m. The Kylylahti massif is reached at 200 metres relative depth, first as soapstones (SS) and serpentinites (SP), then as

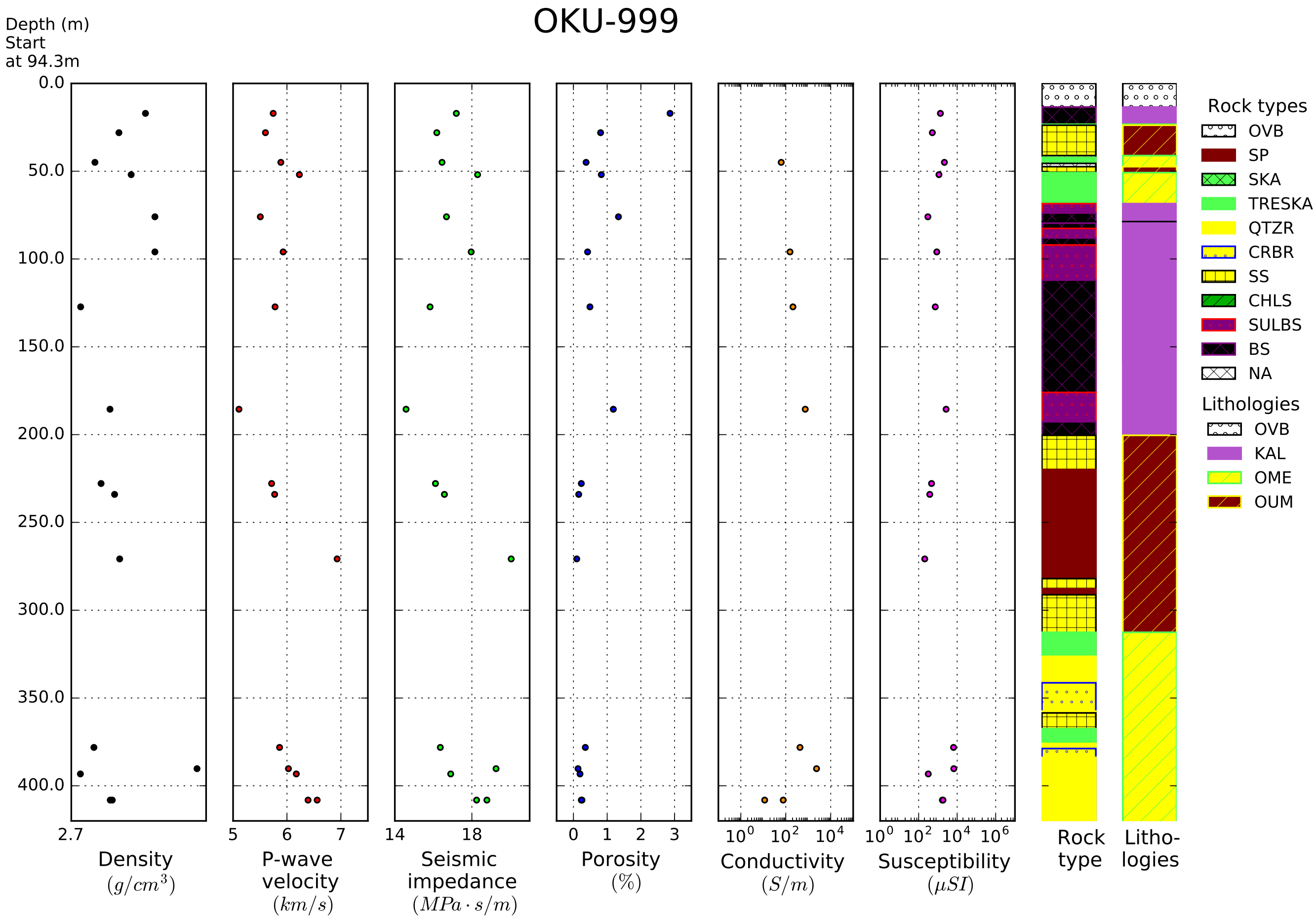

Figure 35: Data for the drill hole OKU-999.

carbonate-skarn-quartz rocks (OME). At its bottom the drill hole reaches the Kalevian rocks anew, at the relative depth of 428 m.

OKU-999 is quite sparsely sampled and the samples mainly serve as complementary samples to their respective rock types. Based on the average seismic impedances for different rock types, the contact between Kylylahti massif and Kalevian rocks surrounding it should be visible in seismic data (Figure 11 and Table 8). The data from OKU-999 does not support this view, which is most probably due to sparse sampling. The data from this drill core indicates, that based on their conductivities, black schists cannot be separated from Outokumpu assemblage rocks containing disseminated sulphides. The highest susceptibility values in OKU-999 are measured for two quartz rock samples containing disseminated ore. From OKU-999 there are so few samples, that no strong conclutions about the petrophysical differences between the rock types can be made solely based on results from it.

The drill hole OKU-1000 (Figure 36) is a bit eastward from OKU-999 starting from the surface at 98.2 m and it is 324.2 m long. It first runs through Kalevian rocks (KAL), then reaches Kylylahti massif at the relative depth 125 m. Next 130 metres are carbonate-skarn-quartz rocks (OME) and the hole returns to Kalevian rocks again at about 260 metres. In this study we have 10 samples from OKU-1000,

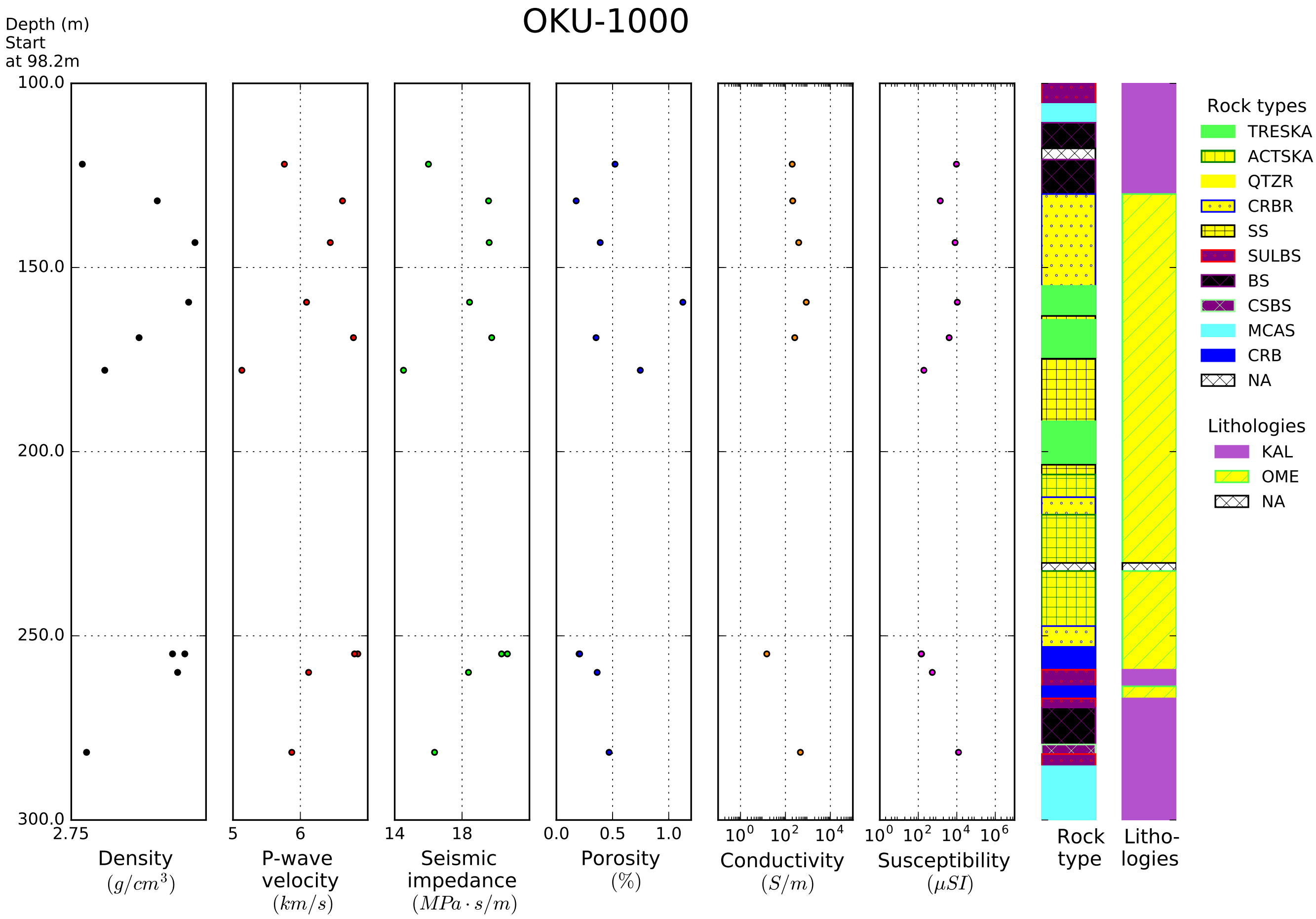

Figure 36: Data for the drill hole OKU-1000.

complementing the other similar rock type samples. The results are quite similar to those from OKU-999, except that OKU-1000 data supports the view that the contact between the Kylylahti massif and the Kalevian rocks (125 m and 260 m) surrounding it produces a detectable reflection in seismic data.

The drill holes OKU-1002 and OKU-1003B are a few hundred metres to the west from the other drill holes mentioned in this study (Figure 29). Both start from the surface. The drill core 1003B represents Kalevian metasediments surrounding the Kylylahti massif. Because there are only five samples of mica schists from a more than 150 metres thick layer of mica schists intervened only by two thin quartz veins, these results are not shown in figure. The petrophysical properties of OKU1003B samples do not deviate remarkably from the average properties of typical Kylylahti region mica schists.

Drill core OKU-1002 (Figure 37), runs first through Kalevian rocks (KAL). It encounters the Kylylahti massif at about 400 metres relative depth, the massif margins here consisting of soap stones (SS), chlorite schists (CHLS) and serpentinites (SP). The ten samples from this drill core are classified as tremolite skarn (TRESKA), sulphide-rich black schist (SULBS), chlorite schist (CHLS), biotite, serpentinite (SP) and soap stone (SS) covering the depth interval 354-455 m from a drill core that starts 115.7 m above the sea level and is in total 513.1 m deep. This drill core is close to the seismic line E1 (Figure 30) discussed

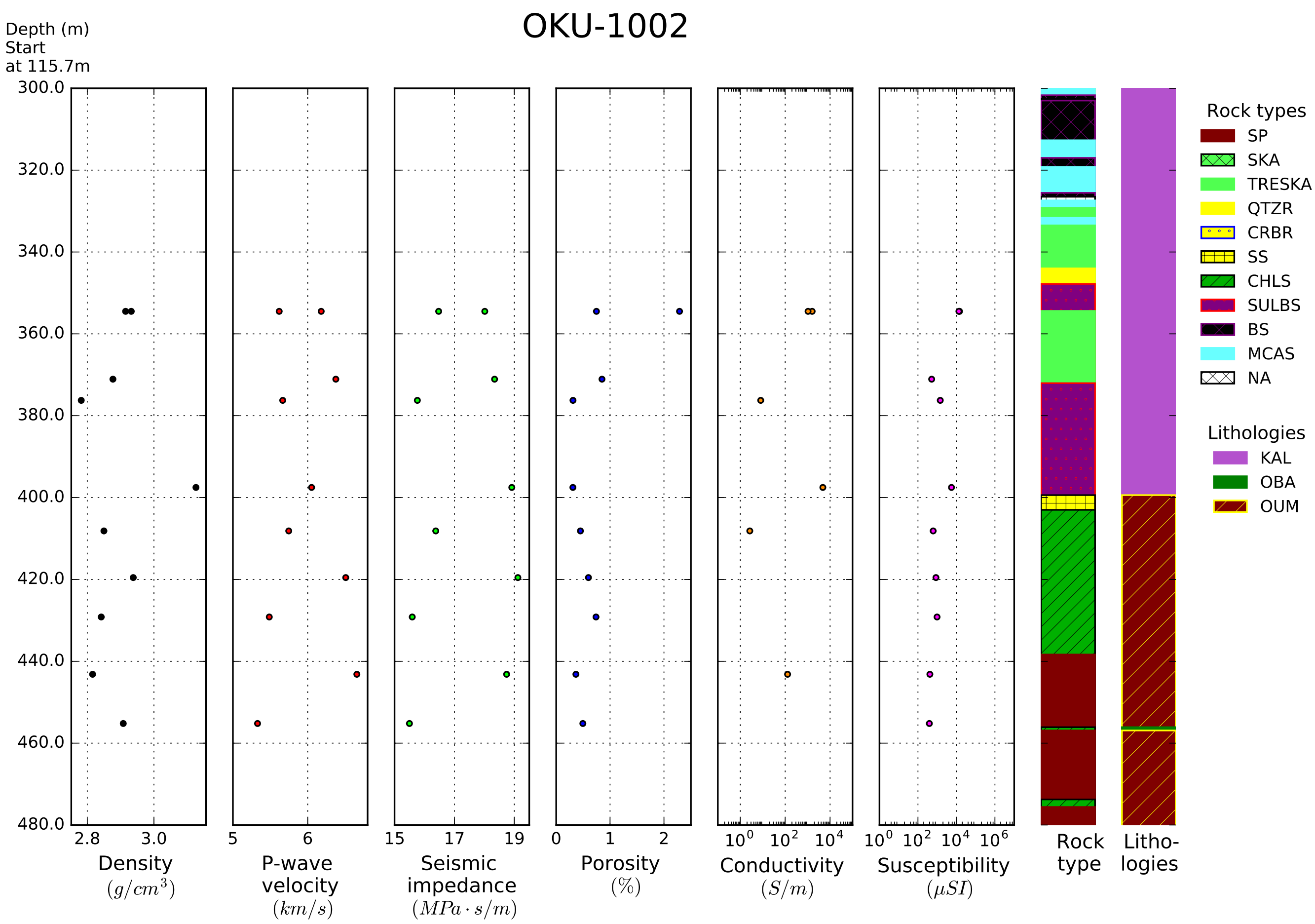

Figure 37: Data for the drill hole OKU-1002.

later (Chapter 6.5).

From the seismic impedance averages (Figure 11) it is clear, that the Kylylahti massif margin should be detectable in seismic reflection data. Data from the drill core OKU-1002 (Figure 37) indicates that there are possible reflectors, but the locations of the reflectors in the drill core log, are not as evident. The mica shists (MCAS) have an average seismic impedance of 15.5 MPas/m (Table 8). Roughly half of the samples from the drill core OKU-1002 have quite similar seismic impedances, whereas the rest of the samples have much higher ones. The problem is, that the value seems to be high and low for every other sample. More samples would be needed to locate reflectors on drill core data.

Otherwise the data from OKU-1002 is as expected from the average values for different rock types. The porosities are low overall. Sulphide-rich black schists (SULBS) are the most conductive ones. Susceptibilities are low, even for serpentinites (SP). This is due to higher degree of alteration, as was already mentioned when discussing the results from KU-901. The presence of talc carbonate rocks and one sample of almost pure biotite confirm that higher degree of alteration is the reason for low susceptibility of serpentinites in this part of the Kylylahti massif.

6 Discussion

6.1 Considerations on petrophysical data

When making direct interpretation of petrophysical laboratory data or using results to interpret data from geophysical surveys, it is important to evaluate, how well the results describe the real properties of rocks. One question is representativity of the sample sets. Another is, how well the laboratory results correspond to in-situ properties. And finally, even a good sample set represents only individual points in rock. However accurate the point-wise petrophysical data is, a single sample cannot be representative for a whole block of rock. On the other hand, geophysical methods handle bedrock as blocks having some average properties, and not always well defined borders to other blocks with different properties. This integrative nature of geophysical surveying makes it reasonable to use average properties of rock types for interpretation.

6.1.1 Representativity of the samples

Representativity of a set of samples is always to be evaluated, when interpreting the results. Especially, when it comes to physical properties of rocks, one has to keep in mind that they are highly variable even within small areas. Meaningful averages for a rock type or a unit can be achieved with large enough number of samples. For example, for the GTK petrophysical database, three samples were taken from each datapoint (outcrop), and an average of the results is used as the result for that point (Airo and Säävuori 2013). Considering this, the sample set used in this study, is not big. Usually there is just one sample per data point. For each rock type, we have 15 samples on average. Well represented are quartz rocks and serpentinites with 43 and 37 samples, respectively. On the contrary, carbonate rocks (non-OKU type), black schists, mica schists, tremolitic calc-silicate rocks and skarns are all analyzed with less than ten samples per rock type. Fortunately, there is a big petrophysical dataset from the Outokumpu district, and from the Kylylahti mine, to which to compare the results of individual studies.

To evaluate the variance of the properties within small areas, three long samples, approximately 30 cm long, were divided into a and b samples. These double samples proved to be quite close to each other in their properties (Table 10). For each pair of A and B samples the difference between the measured values was smaller than within the same rock type generally. An exeption were P-wave velocity and resistivity values for one pair of samples representing tremolitic skarn. Sample A had P-wave velocity of 5621 m/s and galvanic resistivity close to 50000 Ωm. Sample B had velocity of 6182 m/s and resistivity 2250 Ωm. The difference cannot be explained with fracturing. Then the more resistive sample should be less fractured, and have higher P-wave velocity. The reason must be in the mineralogical differences of the samples. This example makes it clear, that there has to be an adequate amount of samples, to make meaningful averages.

Table 10: Petrophysical properties of double samples

Sample	Rock type	Density kg/m^3	Porosity PE(%)	Susceptibility µSI	Remanence mA/m	P-velocity m/s	Resistivity Ωm
999-15a	CRBR	2863	0.253	1704	526	6557	10800
999-15b	CRBR	2855	0.233	1832	209	6395	13200
1000-7a	CRB	3019	0.202	147	63	6852	132
1000-7b	CRB	2990	0.207	141	131	6810	132
1002-1a	TRESKA	2932	2.289	14383	5161	5621	52900
1002-1b	TRESKA	2915	0.747	13406	6075	6182	2250

6.1.2 Effects of pressure, temperature an anisotropy on petrophysical properties

The sampling procedure itself affects the results. Coring and recovery change stress and temperature of the sample, and may change its structure. The sample can be further changed when transporting, cleaning and otherwise handling the samples (Schön 2015). Especially the seismic velocities measured in laboratory are easily affected by microcracking due to decompaction during the core retrieval. The effect is the bigger, the bigger the change in pressure has been, i.e. how deep is the origin of the sample (e.g. Kern et al. 2001). This effect can be compensated, if the velocity is measured under pressure and temperature equivalent to the samples's in-situ conditions, but all the P-wave velocity measurements in this study were conducted in atmospheric pressures. Fortunately there are pressurized P-wave velocity measurements from the Outokumpu Deep Drill Hole core ((Elbra et al. 2011, Kern and Mengel 2011). Those give solid basis for estimating the effect of pressure on P-wave velocities also in Kylylahti.

The variation in the velocity of elastic waves in rocks is caused by differences in the mineralogical composition, effects of cracks, fractures and pores, anisotropy effects, temperature and pressure. In dense, non-fractured, igneous rocks, like the majority of rocks in Kylylahti, the velocity of elastic waves is controlled by mineralogical composition (Schön 2015). Even when the in-situ velocities are mainly controlled by mineralogical differences, the results of laboratory measurements can be misleading if not done and interpreted carefully. The effect of microcracks on seismic wave velocities has already been mentioned several times. Also anisotropy of wave velocities is most notable on low-pressure regime. Velocities increase and anisotropy decreases with increasing pressure. The change is the fastest in the low-pressure range that corresponds to the prevalent pressure on first few kilometres on crust. The fast change is entitled to closure of microcracs and large pores. The change is not similar to all rock types, the rocks with high porosity and high degree of microcracking are affected the most. Part of the effect is inherent to the rocks on the site, but part of it is caused by decompaction of samples during the core retrieval (e.g. Kern et al. 2001, Schön 2015).

Elbra et al. (2011) measured P-wave velocities for a set of Outokumpu Deep Drill Hole samples at rising pressure and temperature. The velocities were found out to increase sharply, 100-200 m/s, within the first circa 40 Mpa. That is equal to distance from the surface to approximately 1.5 km depth. As the porosities of the samples were very low, smaller than 1 %, the increase in velocity was probably due to compaction and closing of microcracks. On the bases of borehole logging data, Elbra et al. (2011) conclude that there are less microcracks in situ than in the samples. That means that an important part of the fracturing is due to decompaction during core retrieval. The velocity dependence on temperature was found to be quite small. The results were somewhat different for different rock types, but comparison is difficult. Most of the samples (98 pieces) represented mica schists and other rock types had only one or at most four samples that were measured under estimated in-situ conditions (Elbra et al. 2011).

The velocity anisotropy of Outokumpu Deep Drill Hole samples has been measured and modelled by Kern and Mengel (2011). They found out that in Outokumpu, there is a strong directional dependence of wave propagation in biotite gneisses. It would enhance the seismic reflectivity in the contact of biotite gneisses and ophiolite-related Outokumpu assemblage rocks in vertical direction, but if the geometry was different, the contrast would not be clear. It is well possible, that also in Kylylahti, anisotropy plays some role in the P-wave velocities and it should be studied more thoroughly.

Considering the effects of pressure on velocities, and on the other hand, the fact that the porosity of Kylylahti rocks is very low overall, it is justified to use the P-wave velocities measured in this study as a reference when interpreting seismic data. Comparing the velocities to the ones measured by acoustic drill hole logging would be beneficial. Probably the velocities measured in this study are somewhat biased towards lower values than the real in-situ velocities due to the pressure effects, especially when deeper parts of the Kylylahti area are conserned.

When it comes to other petrophysical parameters discussed here, the pressure difference between laboratory and in-situ conditions is not significant. Increasing pressure affects the physical properties of rocks first through reduction of porosity. This effect could theoretically increase the resitivity and density of water-saturated rocks, even on pressures below 30 MPa, the approximated maximum pressure at our sampling sites (e.g. Schön 2015). However, no systematic decrease in porosities as a function of depth was found in our data (Figure 14), and thus it can be assumed, that in low-porosity environments, like in Kylylahti, the pressure is not an important factor for densities or electrical properties of rocks at depths shallower than one kilometre. The same applies to magnetic properties. Magnetic susceptibilities are expected to decrease with increasing pressure (Schön 2015), but at the pressure range encountered in this study, mineralogy is expected to dominate the susceptibility differencies. This was also the case in the Outokumpu Deep Drill Hole core (Airo et al. 2011).

Temperature does not have any significant impact on physical properties of rocks at the depths discussed in this study. In the Outokumpu Deep Drill Hole, the temperature was 40 °C at 2.5 km depth (Kukkonen 2011). There is no reason to expect, that the underground temperatures would be nuch higher in Kylylahti, only a bit more than twenty kilometres away from Outokumpu. All the COGITO-

MIN samples are from shallower depths than 2.5 km. The low temperature has practically no impact on any of the petrophysical properties measured in this study (e.g. Schön 2015).

6.1.3 Effect of sample soaking time on the P-wave velocities

One important thing, regarding the reliability of P-wave velocity laboratory measurements, seems to be the soaking time of samples before the measurements. It has been known for a long time, that water saturation partially compensates for the effects of microcracking (e.g. Salisbury et al. 2003, and the references therein). In the GTK petrophysical laboratory P-wave velocities of rock samples are nowadays always measured under water and samples are soaked before the measurements. In previous measurements it has been noticed that measured P-wave velocities increase significally with soaking time (H. Säävuori 2016, pers. comm.). An example is in Figure 38. A set of Outokumpu Deep Drill Hole samples was measured every four days to see how the P-wave velocity changes with increasing soaking time. The results show, that during one month, in the lower velocity group (velocities clearly under 5000 m/s), the velocity increases about 550 m/s and in the higher velocity group about 120 m/s. The same phenomenon has been observed also for another set of drill core samples (H. Säävuori 2016, pers. comm.).

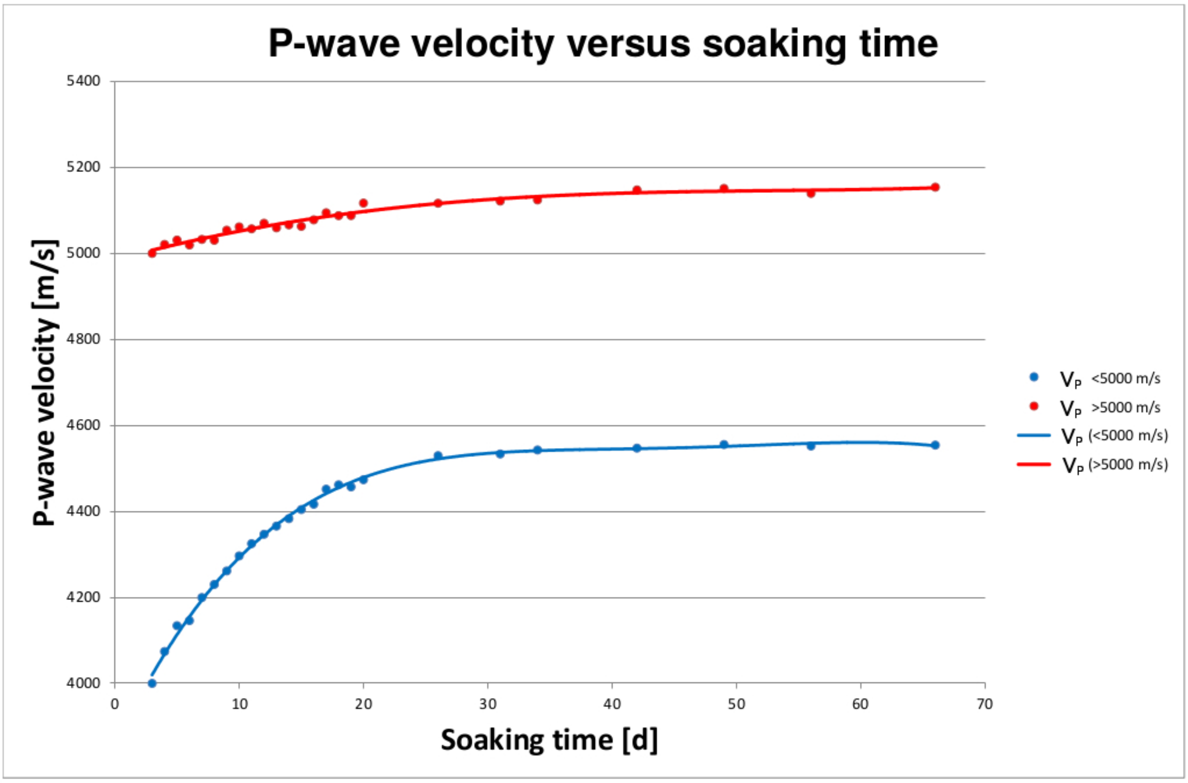

Figure 38: The effect of soaking time to the measured P-wave velocity. The change of velocity is shown in two velocity groups; over 5000 m/s and velocities clearly lower than 5000 m/s. The grouping is done after 4 days soaking times. In the picture each point represents a mean value of six samples. The rocks are from the Outokumpu Deep Drill Hole, mostly mica schists, but there are also other rock types in both sets of samples. In the lower velocity group (blue) the average change has been added to velocity of 4000 m/s and in the higher velocity group (red) the average change has been added to the velocity of 5000 m/s. (Figure: GTK petrophysical laboratory/Heikki Säävuori.)

To get more data on this partially unexplained behaviour, COGITO-MIN samples were measured three times with approximately 10 days intervals. Unfortunately, for detecting this phenomenon, all the samples from Kylylahti had high P-wave velocities and the results were not so evident. Afterwards, it was also realised, that the velocities should have been measured for the first time after a few days, not ten days. The increase in velocity was not clear for every sample, but the average velocity of all samples after ten days was 6135 km/s, after 20 days 6166 km/s, and 6174 km/s after a month. On average, the velocity increased 39 km/s. This supports the previous observation that soaking time does not affect significantly the rocks with high velocities.For low velocity samples, the optimal soaking time should still be studied, as well as the precise mechanism, how saturation increases the velocities. These questions suggests further research on this subject.

6.2 Results of this study compared to other petrophysical results from the Outokumpu district

One of the motivations for this work was to complement the existing petrophysical data from the Outokumpu district, and especially from Kylylahti. Therefore it is essential to compare the new results with the previous data.

6.2.1 Density

In Figure 39 is a density-susceptibility diagram for Outokumpu region outcrop samples from the GTK petrophysical database and for COGITO-MIN Boliden samples set. GTK samples represent the Archean basement (granites, granodiorites, granite gneisses and gabbros), Jatulian quartzites and Kalevian mica schists. Kylylahti samples, mostly Outokumpu assemblage rocks, have notably higher densities than the rocks surrounding them. Especially the sulphide mineralizations are much denser than the other rock types, but also the Kylylahti serpentinites have two population, another with anomalously high susceptibility and density. The entire Outokumpu assemblage differs from its background by its density. This was already noted from the Outokumpu Deep Drill Hole core samples. In Airo et al. (2011), Elbra et al. (2011), and Heinonen et al. (2011), it was noted that the most abundant rock type, mica schist was quite monotonous rock type, having densities between 2700 and 2800 kg/m^3. The greatest variation in density was present in the Outokumpu Assemblage rocks, found between 1300 m and 1500 m in the drill core. Most of these rocks had densities over 2800 kg/m^3. This is in accordance with our results from Kylylahti. However, when analyzing the results for each rock type, there are some notable deviations between the Kylylahti rocks and other Outokumpu district rocks.

Physical properties of rocks in the Outokumpu district reflect the metamorphic zonation (e.g. Säntti et al. 2006, Leväniemi 2016). Densities of serpentinites, shown in Figure 40, are one example of that. The chrysotile serpentinites in the west (Outokumpu, Keretti, Vuonos, Miihkali) are on average notably lighter than the antigorite serpentinites in the east (Horsmanaho, Kylylahti). Even considering this,

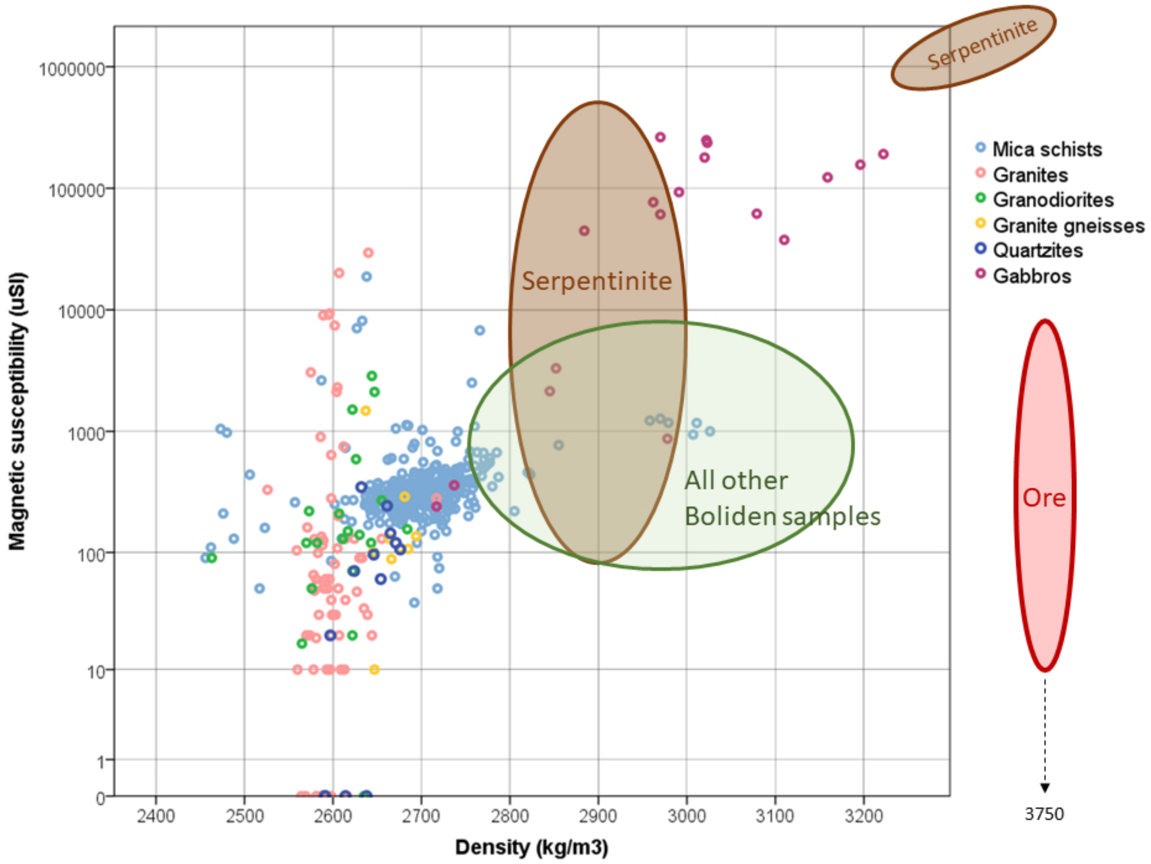

Figure 39: Density-susceptibility diagram for Outokumpu region outcrop samples from the GTK petrophysical database and for COGITO-MIN Boliden samples set. The samples of rock types in the GTK database are plotted according to the legend. The range of density and susceptibility values of Boliden sample set ore samples (MS, SMS), serpentinites (SP, two populations) and other rock types, are marked in the figure by red, brown and green areas, respectively. Modified from (Leväniemi 2016)

the COGITO-MIN serpentinite samples are anomalously dense compared with the average for other serpentinite samples from Kylylahti or elsewhere in the Outokumpu district. However, the median value of COGITO-MIN samples is quite close to the average of other Kylylahti samples. In closer examination, it appeared that all the anomalously dense samples were from one sequence of the drill core KU-901 (Figure 41), which has been classified as an iron formation in Kylylahti drill core database (Boliden 2016b), although the samples themselves were classified as serpentinites. If these samples are left out of the average, the mean density of COGITO-MIN serpentinites is 2828 kg/m^3, which is already much closer to the average of other Kylylahti samples.

An iron formation (IF) in the serpentinites of Kylylahti massif has been described by Kontinen (2005). It was dissected by drill holes OKU-791 and OKU-794. These holes are less than 150 metres from the drill hole KU-901, on horizontal plane, and the depth of iron formation of the drill hole OKU-794, and of serpentinites in the drill core OKU-791, roughly coincides with the depth of the drill core KU-901 iron formation. In this formation veins or patches in serpentinites have been replaced. The replacements are magnetite and pyrrhotite rich (Kontinen 2005). Apparently the samples from the KU-901 IF-zone belong to that category, as they also have high susceptibilities (Figure 40).

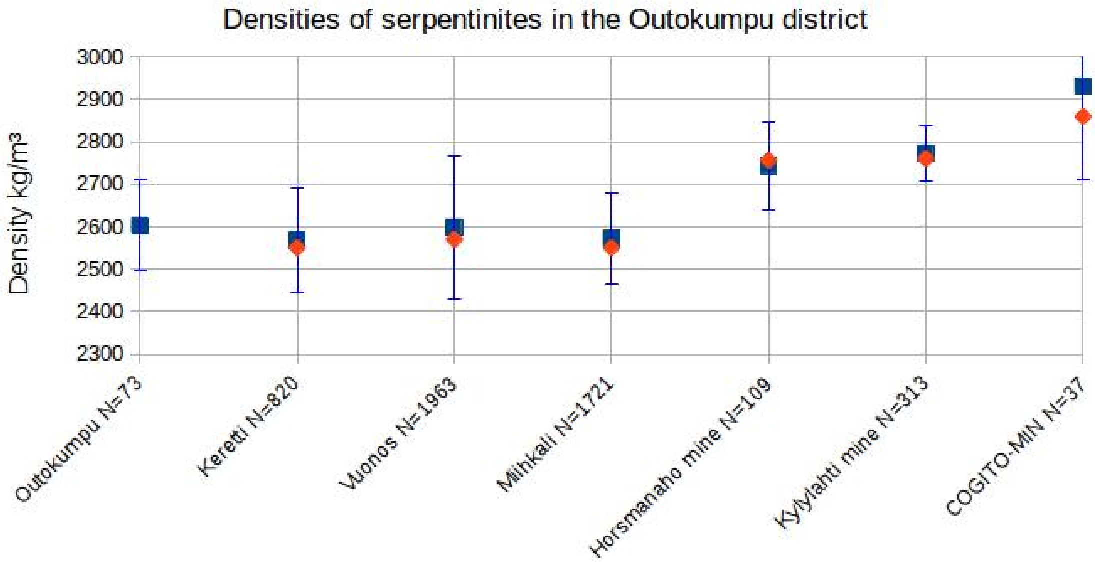

Figure 40: The average densities (brown boxes) and medians (orange dots) of serpentinite samples from drill core samples in the Outokumpu district. Sampling sites are ordered from right to left according to their whereabouts from east to west. Standard deviations are shown as error bars. Values are from (Leväniemi 2016) except for Outokumpu (Heinonen et al. 2011) and COGITO-MIN (this study). Both Kylylahti and COGITO-MIN are samples from Kylylahti. Kylylahti contains density values from the Kylylahti mine drill core database (Boliden 2016b), COGITO-MIN values are measured in this study. Number of samples from each site is marked after the name of the site.

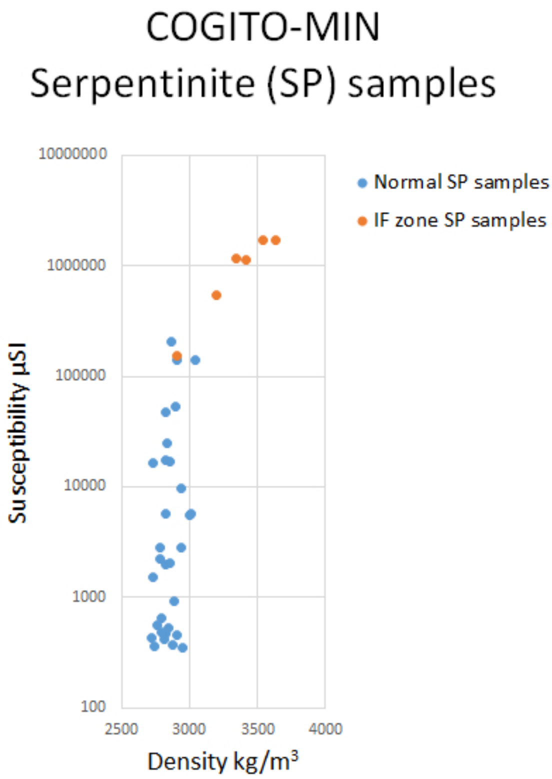

Figure 41: Susceptibility-density diagram for COGITO-MIN serpentinite samples. Normal samples are from the lithological units logged as OUM (OKU serpentinites and soap stones) and IF zone samples serpentinite samples from iron formation unit.

Soap stones, talc-carbonate rocks, are found on altered margins of ultramafic bodies in the eastern Outokumpu district, like in Kylylahti and Horsmanaho. In the western parts, the metamorphic grade is too high for soap stones to exist. The average density of the COGITO-MIN soap stone samples (2900 kg/m^3) is practically the same as the average of Horsmanaho mine soap stones (2901 kg/m^3), reported by Leväniemi (2016). Leväniemi (2016) reports 2811 kg/m^3 for the Kylylahti mine soap stones. The difference between Horsmanaho and Kylylahti may be explained by different rock-type classification. The change from serpentinites to soap stones is gradational Kontinen (2005) and the exact limit between the two rock types is a matter of convention. It is possible, that the convention for classification is different in these two mines, especially when soap stone is the rock mined in Horsmanaho. The COGITO-MIN soap stone samples have been chosen to be of as pure soap stone as possible whereas the entire Kylylahti mine database (Boliden 2016b) may contain also less pure soap stones. Serpentinites are less dense than soap stones, even in Kylylahti when the IF-zone serpentinites have been excluded, and thus less pure soap stones are lighter than the pure ones.

In her analysis, Leväniemi (2016) has handled carbonate rocks, skarns and quartz rocks as one rock type, called 'other Outokumpu assemblage rocks of ultramafic origin. These carbonate-skarn rocks are discussed together in several other contexts too. This is geologically reasonable, because they occur in the fringes of the serpentinite massifs and are very much mixed and folded with each other. These rocks also have quite similar densities. This rock group contains disseminated sulphides and the amount of sulphides affects the petrophysical properties of a sample; densities as well as elastic, electrical and magnetic properties. Therefore this rock group shows a wide range of values for any of these properties, although the averages for each rock type are quite similar. The average density for carbonate-skarn-quartz rocks in COGITO-MIN samples is 2928 kg/m^3. Leväniemi (2016) reports an average density of 2926 kg/m^3 for the other Outokumpu assemblage rocks of ultramafic origin from Kylylahti mine, calculated with densities for 7802 samples.

Average densities for Outokumpu district mica schist and black schist are shown in Figure 42. The values reported for drill core samples are somewhat higher than for the outcrop samples from the GTK petrophysical database, except for the Outokumpu Deep Drill Hole core. The amount of mica schist samples is big for each sampling site, whereas the COGITO-MIN sample set is small, containing six samples of mica schists. From Kylylahti and Horsmanaho mines there are around 250 density values, more in other data sets and the most from Outokumpu Deep Drill Hole, 7929 values. Big amount of samples means that these averages are well representative of their rock type. Additionally, standard deviation is small for mica schists.

According to Kontinen (cited by Leväniemi (2016)), the higher mean density of mica schists in drill core samples is probably due to two reasons. First, the drill core samples, except for the Outokumpu

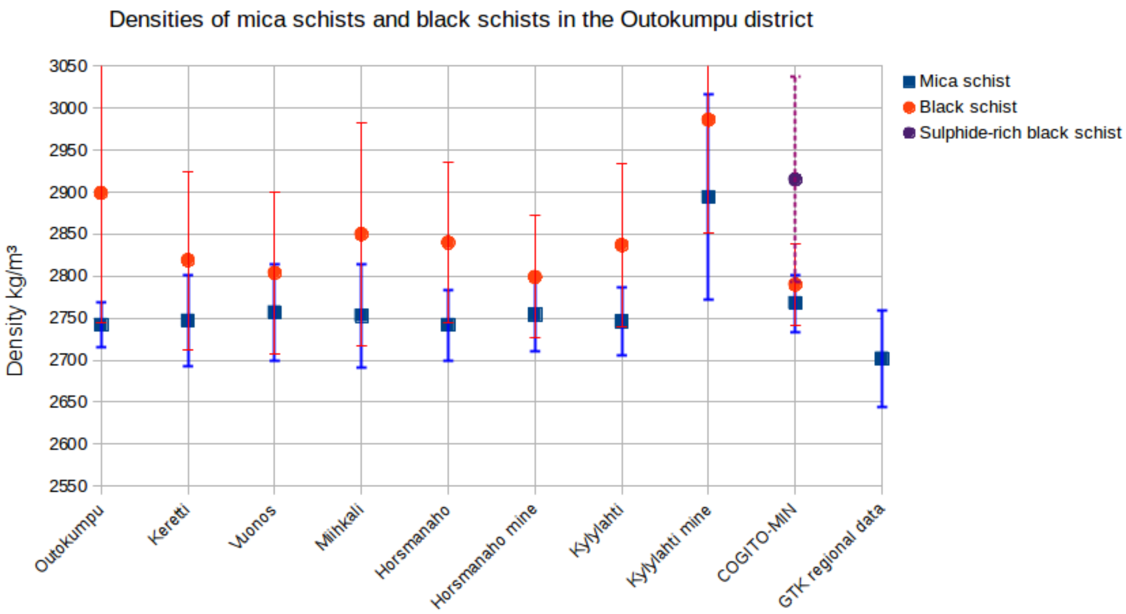

Figure 42: Average densities for mica schists and black schists from drill core samples in the Outokumpu district. Sampling sites are ordered from right to left according to their locations from east to west. GTK regional data is from the outcrops all over the Outokumpu district, outside of the mining sites. Standard deviations are shown as error bars. Values are from (Leväniemi 2016) except for Outokumpu (Airo et al. 2011) and COGITO-MIN (Kylylahti) (this study).

Deep Drill Hole core, are mostly from the environments rich in sulphide mineralizations. There the mica schists often occur close and grade into sulphidic black schists. Second, the GTK petrophysical database samples are outcrop samples and thus are naturally more weathered than the drill core samples, their higher porosity decreasing the density. The mica schists from the Kylylahti mine database (Boliden 2016b) have much higher average densities than the other mine environment mica schists. On the contrary COGITO-MIN mica schist samples fall nicely to the same category with the others. The question of the average density of mica schists, surrounding the Kylylahti massif from all the sides, is of course important for seimic studies. In general it is safe to assume, that there is a big contrast between the mica schists and Outokumpu assemblage rocks. Nevertheless, the reason for high densities of mica schists in the Kylylahti mine database remains unsolved and suggests more research on that subject.

The densities of black schists, also shown in Figure 42, behave the same way as the densities of mica schists. Otherwise averages are quite uniform but Outokumpu Deep Drill Hole and Kylylahti mine have higher value. In COGITO-MIN samples suphide-rich black schist has been classified as its own rock type and thus the densities of black schists (not rich in suphides) are quite low. From other sites it remains unclear, how many samples could have been classified as sulphide-rich. For black schists, the standard deviation of averages is higher than for mica schists, because there is so much variability in the sulphide content of the samples.

For the sulphide mineralizations there is not much density information from the Outokumpu district except for Kylylahti. The mean density for the ore samples, massive and semi-massive, in the Kylylahti mine database (Boliden 2016b) is 3454 kg/m^3. The standard deviation of ore density values is large, 230 kg/m^3. According to Tuomi (2016), the density increase clearly correlates with the increase in sulfur and iron content. In the COGITO-MIN-samples the mean density was 3803 kg/m^3 for massive and 3705 kg/m^3 for semi-massive ore. Considering there is almost 900 density values in the mine database and only 34 ore samples in COGITO-MIN set, it can be stated that the average density of sulphide mineralizations is closer to 3500 kg/m^3 than 3700 kg/m^3. However, sulphide mineralizations are much denser than the rocks surrounding them, except for the rocks with ample sulphide disseminations.

6.2.2 P-wave velocity

P-wave velocity data from the Outokumpu district was restricted to Outokumpu Deep Drill Hole before this work. Petrophysical properties of the Deep Drill Hole core samples were reported by Airo et al. (2011). P-wave velocities were discussed in more detail by Elbra et al. (2007b) and Elbra et al. (2007a). Not all the Kylylahti rock types were present in the Deep Drill Core. Notably, no sulphide mineralizations were found. Heinonen et al. (2011) reports P-wave velocities measured by downhole logging. Elbra et al. (2011) made laboratory measurements both in atmospheric and approximated in-situ pressures. The sample set of Elbra et al. (2011) had only mica schist, black schist and serpentinite common rock types with the sample set of this work. Elbra et al. (2011) pointed out that there is a large variation in seismic velocities, both when measured in ambient pressure and under in-situ crustal conditions. Recently, Tuomi (2016) estimated parameters for seismic forward modelling at Kylylahti based on existing petrophysical data from the Outokumpu district. These studies coupled with the present study, will be used to discuss recommended parameters for seismic modelling (Chapter 6.3).

In the results of this work (Figure 11), we have two clear categories in P-wave velocities: the low velocity rocks (mica schists, black schists and soap stones) having velocities close to 5.6 km/s and high velocity rocks (Outokumpu assemblage carbonate-skarn-quartz rocks, Kalevian carbonate rocks and sulphide mineralizations) having velocities well over 6 km/s. Sulphide-rich black schist forms an intermediate group between these two groups. Velocity variation inside the rock types is large, for example for massive ore from 5.4 to 7.1 km/s. In the Outokumpu Deep Drill Hole data and parameters chosen by Tuomi (2016), this division is not as clear. The differences will be discussiden in detail in Chapter 6.3

6.2.3 Porosity

The only reference to porosities for this study is from the Outokumpu Deep Drill Hole core. Airo et al. (2011) report that the porosities are very low overall. Interestingly, the porosity of mica schists increases with depth, but the densities remain the same. It is thus expected to be due to mineralogical changes in the mica schists. Otherwise, the porosity results are consistent with COGITO-MIN sample porosities.

6.2.4 Magnetic properties

Figure 39 shows magnetic susceptibilities of rocks in the Outokumpu district. Mostly the rocks are in the paramagnetic population. Exceptions are ferrimagnetic serpentinites and gabbros that are known to be magnetite-rich (e.g. Airo et al. 2011). Magnetic susceptibility values are quite difficult to compare, due to their high variability. The values reported by Leväniemi (2016) for Kylylahti mine are in line with the results of this work for all rock types. Serpentinites have higher susceptibility values in the western Outokumpu district than in the eastern. This is readily explained by differences in metamorphic grade. The chrysotile serpentinites of higher metamorphic grade in the west are expected to have higher susceptibility values than the lower metamorpic grade antigorite serpentinites in the east (Leväniemi 2016). The black schist susceptibilities also decrease towards the east. It can be due to decrease in the amount of pyrrhotite or the presence of hexagonal, antiferromagnetic pyrrhotite, which has been detected in the black schists of the Outokumpu region (Airo et al. 2011, Västi 2011).

There is less data on the intensities of remanent magnetization of the Outokumpu district rocks, than there is susceptibility data from the area. Consequently, Q ratios can be calculated only for a few sample sets. Leväniemi (2016) has found two references for her study. A sample set of black schists from the Outokumpu district outcrops, measured by Outokumpu Oy, has Q ratios close to ten. Horsmanaho mine data includes also measured intensities of remanence. From these results it is confirmed, that especially for black schists, remanence is an important factor as a magnetic anomaly source.

Outokumpu Deep Drill Hole core samples have been extensively investigated, and also remanent magnetization has been measured. In Outokumpu, the average remanence is weak along the whole Deep Drill Hole core. Only the Outokumpu assemblage rocks on the core have notably high intensities of the remanent magnetization. The average Q ratios for different rock types are under one, except for chlorite muscovite schists and biotite gneisses, and only individual samples have Q ratios much higher than one (Airo et al. 2011). This is very similar to our results from Kylylahti. However, high intensities of remanence in part of the rocks, and especially in the outcrop black schists, requires consideration when interpreting magnetic anomalies.

6.2.5 Electrical properties

Airo et al. (2011) measured also specific resistivity of the Outokumpu Deep Drill Hole core samples. Majority of the samples were silicic, so called common rocks, having resistivities between 1000 and 100000 Ωm. The sequence of Outokumpu assemblage rocks contained more conductive samples. This was entitled to them containing sulphides and graphite. Within the mica schists the resistivities decreased with depth, whereas the porosities increased. This was supposed to be partly due to some, not yet specified, mineralogical changes.

The electrical properties of the Outokumpu district rock types reported in this work, or by Airo et al. (2011), support the conclusions done by Airo and Loukola-Ruskeeniemi (2004). According to them, Outokumpu-type sulphide mineralizations are moderately conductive. This is due to sulphide mineralizations but also the black schists commonly related to them. The zones most rich in sulphide mineralizations may even appear less conductive than the area immediately around them. According to our results (Figure 24) massive and semi-massive ore are clearly more conductive than the other Kylylahti rocks, but many other Outokumpu assemblage rock types have also individual samples with high conductivity and they are overall slightly conductive. Blacks schists were clearly less conductive than the ore, but they were only represented by three samples. All this point out that the Outokumpu-type sulphide mineralizations cause electrical, conductive anomalies, but these are hard or even impossible to distinguish from the anomalies caused by other conductive Outokumpu assemblage rocks or black schists.

6.3 Parameters for seismic modelling in Kylylahti

Based on the results from the laboratory measurements and geology of Kylylahti, the rock clustering and values presented in Table 11 are recommended for seismic modelling of Kylylahti area. If possible, more detailed density model from drill core data is naturally preferred over these average values. For P-wave velocities, no large data set exists. P-wave velocities are usually dependent on the direction of wave propagation and measurements are much less accurate than for density. All this supports using average values for velocity model.

Clustering carbonate-skarn-quartz rocks together is justified by their relatively similar physical properties, but also by their way of occurring together, mixed and grading into each other. On the other hand, the ultra-mafic rocks, serpentinites and soap stones, should not be handled as one unit. There is a considerable difference in their seismic properties and in Kylylahti, soap stones are abundant. Sulphide-rich black schist could be separated to its own category, although the line between common and sulphide-rich black schist is hard to draw when determining the rock types and doing geological models.

The velocity values recommended in Table 11 are next discussed with the previous recommendations given in Tuomi (2016). The recommeded density averages are average values for each unit, measured in this work. The density value of serpentinites is calculated without the IF zone serpentinites as discussed in Chapter 6.2.1.

For the sulphide mineralizations Tuomi gave the P-wave velocity 5.9 km/s, but suggested also a high velocity choice of 8.2 km/s. These velocities were based on theoretical calculations with different equations and assumptions of the mineralogical composition of the ore. According to the results of this

Table 11: Recommended P-wave velocities and densities to be used when modelling Kylylahti rocks

Rock unit	Rock types included	Avg V_P km/s	Avg density kg/m^3
Suphide mineralizations	massive ore semi-massive ore	6.2	3750
Serpentinites	serpentinite	6.2	2830
Soap stones	soap stone	5.5	2900
Carbonate-skarn-quartz rocks	Skarn, tremolite skarn quartz rock, carbonate rock chlorite schist	6.3	2990
Kalevian rocks	black schist mica schist	5.6	2780
Sulphide-rich black schist	sulphide-rich black schist	5.8	2920

work, the lower value is closer to the truth but a bit low. The mean P-wave velocity for the ore samples from Kylylahti in our data is 6.2km/s.

Serpentinites have been given as low velocity as 5.3 km/s by Tuomi (2016). It is higher than the one measured from the Outokumpu Deep Drill Hole by downhole logging (4.6km/s, (Heinonen et al. 2011)) and the same as measured in the laboratory for the Deep Drill core serpentinites (Elbra et al. 2011). Antigorite serpentinites of Kylylahti are denser than chrysotile serpentinites in Outokumpu and thus the P-wave velocity is expected to be higher in Kylylahti than in Outokumpu. If it is as high as our measurements imply, 6.2km/s, it is not different from other Outokumpu assemblage rocks, the sulphide mineralizations included, but it would be much higher than the velocities for mica and black schists.

In her analysis Tuomi (2016) handled separately skarns and quartz rocks, giving them velocities 6.8 km/s and 6.4 km/s, respectively. She bundled carbonate rocks, soap stones, and chlorite schists together giving them the velocity value of 6.8 km/s. According to our results, soap stones should be considered as their own group. They have much lower P-wave velocity, 5.5 km/s, than other Outokumpu assemblage rocks and are quite abundant on the margins of the Kylylahti massif. On the other hand, carbonate-skarn-quartz rocks could be handled as one group, as has been done in some studies (e.g. Leväniemi 2016). This is justified both by them occurring geologically close and mixed to each other, and by their pretty similar petrophysical properties. According to results of our work, this group could have P-wave velocity of 6.3 km/s.

Tuomi (2016) has chosen P-wave velocity 5.4 km/s for mica schists. The average from the Outokumpu Deep Drill Hole is 5.5 km/s from both drill hole logging and laboratory measurements (Heinonen et al. 2011, Elbra et al. 2011). We have already stated that the density of mica schists seems to be somewhat higher in Kylylahti, at least close to the massif, than in the Outokumpu Deep Drill Hole core. Thus, 5.6 km/s, the average P-wave velocity for mica schists in this work, seems to be reasonable. For black schists Tuomi chose 5.75 km/s. Our values, 5.8 km/s for sulphide-rich black schists and 5.6 km/s for black schists are quite close. The Deep Drill Core results are 5.8km/s in-situ (Heinonen et al. 2011) and

5.4 km/s in laboratory (Elbra et al. 2011). With black schists it has to be noted, that there is large variability in all petrophysical parameters, due to variable sulphide content. If sulphide-rich black schists are separated as their own rock type, they must have higher densities and P-wave velocities than less sulphide-rich black schists. Otherwise, some intermediate values need to be chosen. When modelling close to sulphide mineralization it can even be justified to use the sulphide-rich black schist values only.

6.3.1 Seismic impedance and reflection coefficients

When seismic impedance values for Kylylahti rock types (Table 8) and reflection coefficients between the rock types (Table 9)are compared to the values previously used in forward seismic modelling of Kylylahti sulphide deposit (Komminaho et al. 2016, Tuomi 2016), there are notable differences. Partly these differences are due to densities of Kylylahti rocks, that differ from the more western Outokumpu rocks, as has already been discussed. Partly, because previously there has been no P-wave velocity data from Kylylahti. Mica schists and serpentinites have been given lower seismic impedances than should be given, based on our results. Komminaho et al. gave serpentinites 12 MPa·s·m^{-1} and mica schists 14 MPa·s·m^{-1}, and Tuomi 14.86 MPa·s·m^{-1} and 14.58 MPa·s·m^{-1}, respectively. Our values are 18.2 MPa·s·m^{-1} for serpentinites and 15.5 MPa·s·m^{-1} for mica schists. Kommminaho et al. used 16 MPa·s·m^{-1} for soap stones, which is close to our 15.9 MPa·s·m^{-1}, but Tuomi included soap stones to a rock cluster with seismic impedance of 20.4 MPa·s·m^{-1}.

Reflectivity of a lithological contact is described by its reflection coefficient. Anything over 0.06 is considered high enough to produce a detectable reflection (e.g. Salisbury et al. 1996). Elbra et al. (2011) reported strong and distinct reflections visible in FIRE data from Outokumpu caused by diopside-tremolite-skarn in contact with serpentinite, mica schist, and black schist or tremolitic skarn, with reflection coefficients 1.19, 0.126, or 0.112, respectively. According to our data, tremolitic skarn in contact with with serpentinite has reflection coefficient 0.03, with mica schist 0.11 and with black schist 0.10. Clearly reflectivity properties of lithological contacts are not same throughout the Outokumpu district.

In previous studies, (e.g. Komminaho et al. (2016)) it has been expected, that the sulphide mineralizations are reflective against any background. According to our data, this is the case. On the other hand, the difference in seismic impedances between the serpentinites and the other Outokumpu assemblage rocks is much less than has been previously expected. These contacts are thus probably not visible in seismic data. Also the contrast between the surrounding mica schists and the black schists with the Outokumpu assemblage rocks seems to be smaller than expected previously. Mostly it is still big enough to produce a detectable reflection, but in case of abundant sulphides in black schists, the contrast can be too small for a detectable reflection.

6.4 P-wave velocities of sulphide mineralizations and the contrast between them and their host rocks

As has already been discussed, the P-wave velocity of the sulphide minaralization in Kylylahti is quite close to the velocity of the other Outokumpu assemblage rocks, but because the mineralizations are much denser than the rocks surrounding them, there is a significant difference in seismic impedances. The same applies to sulphide mineralizations in general. Figure 43 shows P-wave velocities and densities for most common chrystalline rocks and sulphide mineralizations. It can be seen that sulphide mineralizations do not usually have much contrast with their surroundings in P-wave velocity, but they do have much higher densities and thus also higher seismic impedances. This has been stated in several studies (e.g. Salisbury et al. 1996).

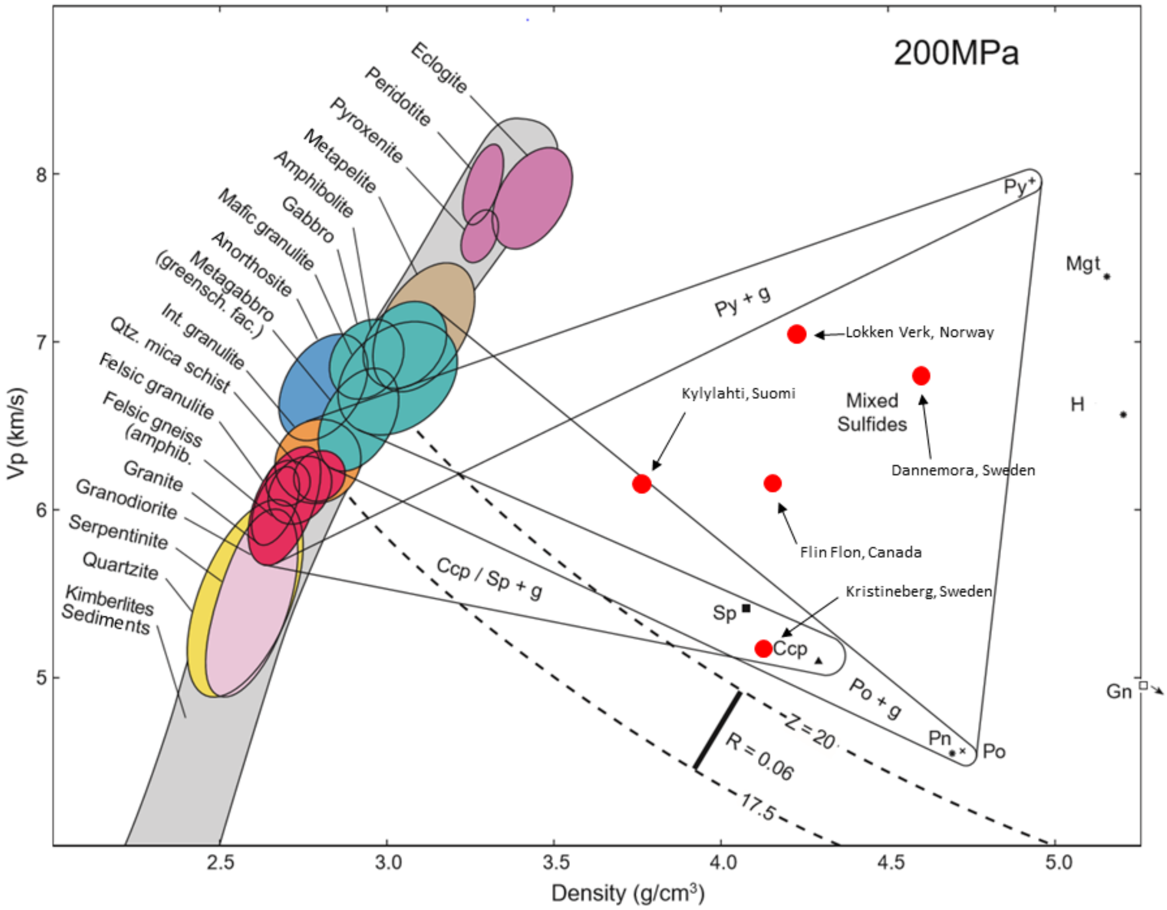

Figure 43: Lines of constant acoustic impedance (Z) superimposed on velocity density fields and Nafe-Drake curve (grey) for common rocks at standard confining pressure of 200 MPa. Also shown are values for pyrite (Py), pentlandite (Pn), pyrrhotite (Po), chalcopyrite (Ccp), sphalerite (Sp), hematite (H), magnetite (Mgt), and fields for host rock-ore mixtures (g mean gang). A reflection coefficient (R) of 0.06 is sufficient to give a strong reflection. Galena (Gn, off scale) has a velocity of 3.7 km/s and a density of 7.5 g/cm^2. Modiefied after (Salisbury and Snyder 2007). Values for magnetite and massive sulphide mineralizations from a few deposits from (Malehmir et al. 2012) and for Kylylahti, this work. P-wave velocities for sulphide mineralizations have not been measured under 200MPa pressure. They would be somewhat higher, if measured under pressure.

Different sulphide minerals have large differences in their velocity and density values (Figure 43). It has been shown (e.g. Salisbury et al. 2000), that velocity-density field of sulphides is controlled by properties of sulphide minerals present and their relative amounts. Pyrite has high P-wave velocity compared to pyrrhotite, thus there should be a difference between the seismic velocities of ore samples from pyrite dominated Kylylahti and pyrrhotite dominated Vuonos. Results of this study support this (Figure 9), also magnetic results(Figures 17 and 19).

It has been stated in several studies (e.g. Salisbury et al. 1996, Salisbury et al. 2003) that large massive suphide deposits should be detectable as reflectors or scatterers. Massive sulphides composed of any mix will make a strong reflector against any likely silicate host except strongly metamorphosed mafic rocks (Salisbury et al. 1996). Metamorphic grade has strong influence on physical properties of rocks, also on reflectivity of a sulphide mineralization. P-wave velocities get higher with increasing metamorphic grade (Salisbury et al. 2003). Also the anisotropy of rocks increases in metamorphic processes. In strongly metamorphosed rocks anisotropy in P-wave velocities can be up to 20%, thus the velocity can be up to 1.9km faster parallel to foliation than it is normal to foliation in schists. If foliation is tectonically induced, it can cause strong reflection even in the absence of any compositional change (Salisbury et al. 1996).

Realistic velocity models are needed for reasonably fast and accurate seismic modelling. The velocities can differ greatly within quite small areas, as can be seen from the velocities measured in the Outokumpu district. On the other hand, if the relationship between metamorphic grade and seismic velocities was better understood, it might be possible to deduce realistic velocities based on geological knowledge of the study area.

6.5 Application of measured seismic impedances in interpretation of seismic data from Kylylahti

As has been stated in several studies (e.g. Malehmir et al. 2012, and the references therein), large massive sulphide deposits are detectable as reflectors. When discussing smaller deposits, the seismic response will be a combination of reflections and diffractions and the shape of the deposit dictates, how clearly it can be detected (e.g Eaton et al. 2003). The results of this study indicate, that in Kylylahti both the ultra-mafic massif and the sulphide mineralizations could be detectable in reflection data, depending on the geometry of the deposit. The seismic forward modelling, done by Komminaho et al. (2016), showed that the near vertical structures found in Kylylahti, are problematic for the interpretation of surface seismic data. On the other hand, borehole data aquisition has shown great potential.

Next a few seismic sections, running close to the COGITO-MIN petrophysics drill holes, are presented. Reflections explained with seismic impedance data aquired from this work, will be discussed. The examples are from the line E1 (Figure 30) and from the in-mine VSP measurements. E1 is a line measured during the HIRE project and the details of the data aquisition and processing are described by Kukkonen et al. (2012a). VSP measurements were a part of the COGITO-MIN project. They were conducted during the summer 2016 and have been shortly described by Riedel et al. (2017b).

Figure 44 shows the line E1 from the southwest. Drill holes OKU-1002 and OKU-1003B, that are close to the line, punctuate the western side of the Kylylahti massif. The upper part of the drill hole OKU-1002 is mica schists, then the rock type changes to diverse Outokumpu assemblage rocks. According to the reflection coefficients, presented in Table 9, the contact between mica schists and any Outokumpu assemblage rocks, except for talc carbonate rock or black schist, produces a strong reflection. And indeed, in the reflection data, there is a strong reflection at the border of the massif. The first, strong reflection is probably caused by the contact between mica schist and tremolitic skarn, having a reflection coefficient 0.11 acoording to our measurements. That contact occurs at approximately 330 metres depth in the drill hole. It is preceeded by a sequence of alterating black schists and mica schists, the contacts of which have reflection coefficient zero.

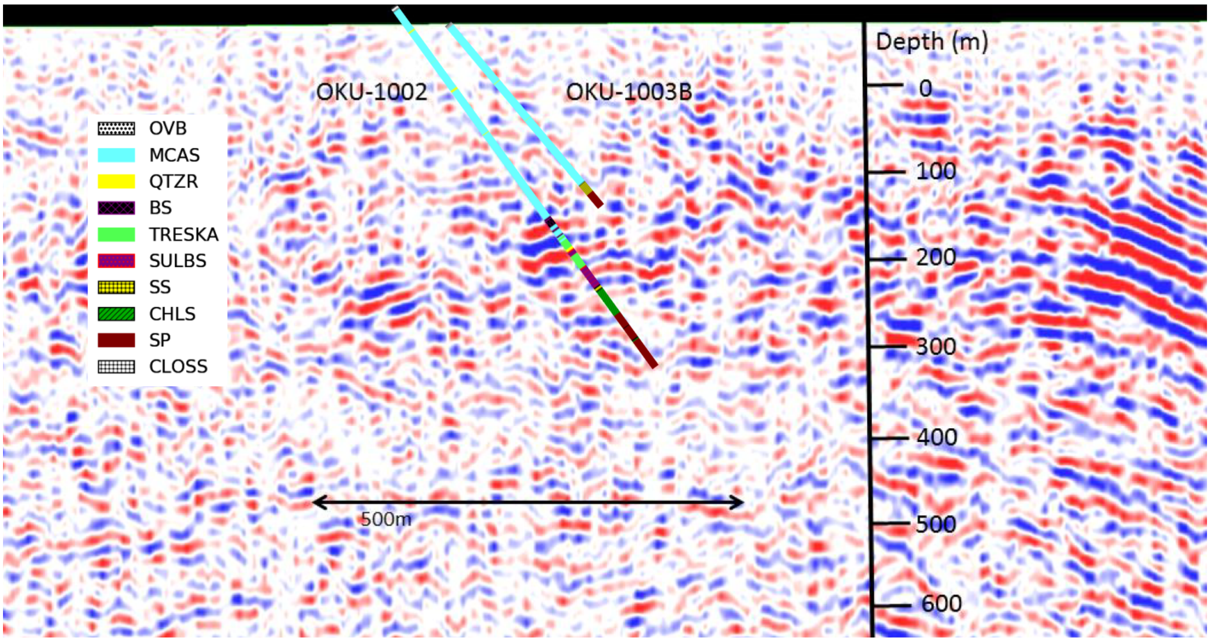

Figure 44: Seismic section E1 seen towards northeast at the location of COGITO-MIN petrophysics drill holes OKU-1002 and OKU-1003B. Legend tells the rock types logged for the drill holes. Rock-type abbreviations can be found from Table 4.

In Figure 45 the seismic section E1 is seen from the northeast. The upper drill holes OKU-999 and OKU-1000 start with Kalevian rocks, mainly black schists this time, and the reach the contact between the massif and its surroundings. The first reflection is probably caused by the contact between soap stone and serpentinite, seen in the drill core OKU-999 (Figure 35, 220m relative depth), having a reflection coefficient 0.07. According to our results, the contact between the Kalevian rocks and soap stones does not produce a detectable reflection. When looking at the lithology of both OKU-999 and OKU-1000 it is clear that there are several contacts that can cause reflections as the mica schists, black schists and soap stones alternate with Outokumpu assemblage rocks with higher seismic impedances. That also explains the polarity change in the middle of the reflective package. It could be caused by the contact between serpentinite and soap stone, having a negative reflection coeffiecient, when waves arrive from the direction of the serpentinite to the soap stone.

On the lower parts of the profile (Figure 45), a reflection from the Kylylahti deposit should be seen, if it is detectable. The deposit is encountered by drill holes KU-901 and KU-900 (Figure 30). No clear reflection can be seen there. On the other hand, there is definitely something, a multitude of strong, but discontinuous reflections. By forward modelling, Komminaho et al. (2016) have shown that these patterns are a hint of vertical reflectors on the area.

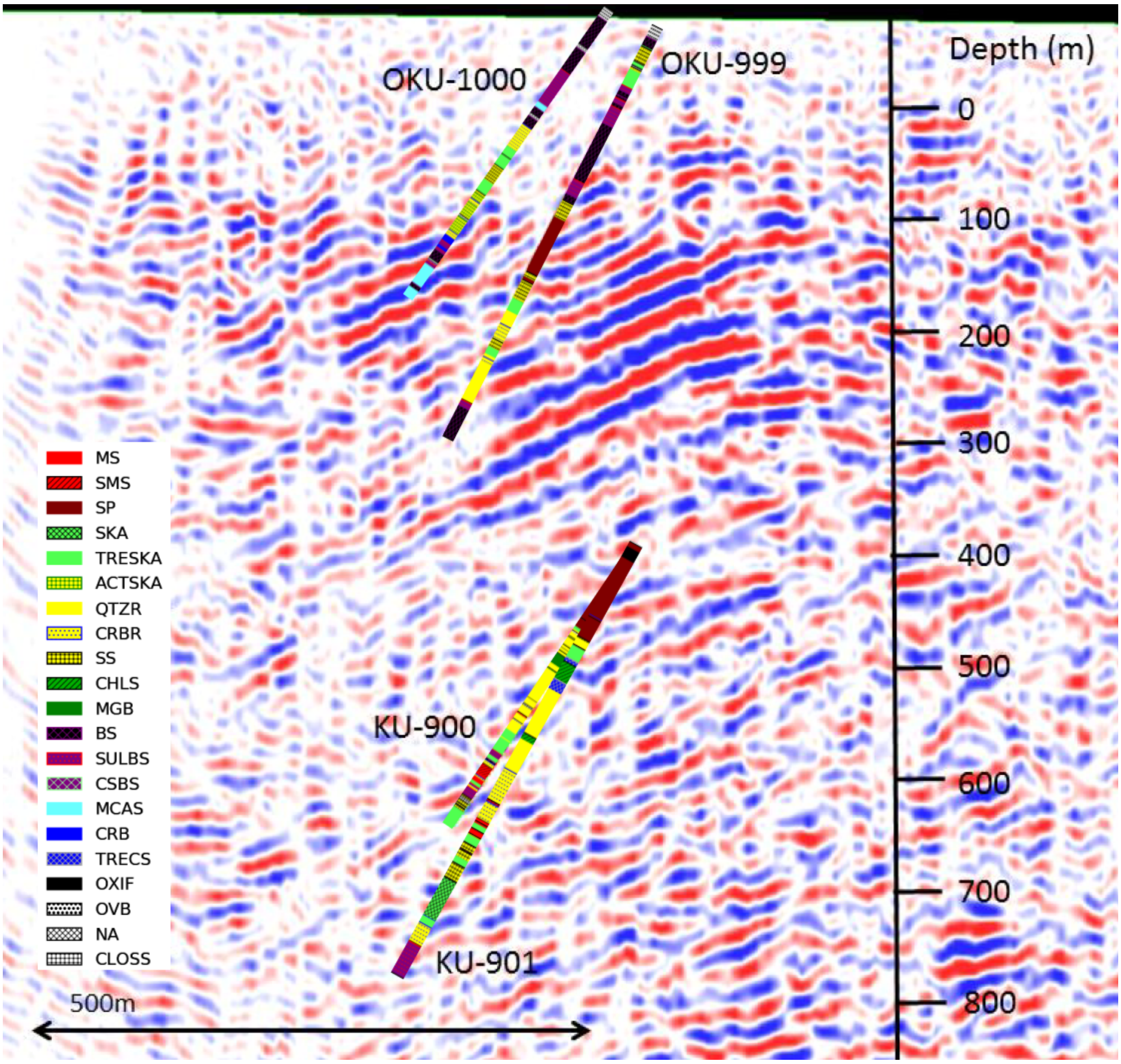

Figure 45: Seismic section E1 seen towards southwest at the location of COGITO-MIN petrophysics drill holes OKU-1000, OKU-999, KU-901 and KU-900. Legend tells the rock types logged for the drill holes. Rock-type abbreviations can be found from Table 4.

In their study, Komminaho et al. (2016) stated that Kylylahti reflectors could be better detected from borehole than surface seismic data. Borehole measurements were a part of the COGITO-MIN field work. Figure 46 shows an example of the preliminary results from COGITO-MIN in-mine VSP data. The data has been reported by Riedel et al. (2018).

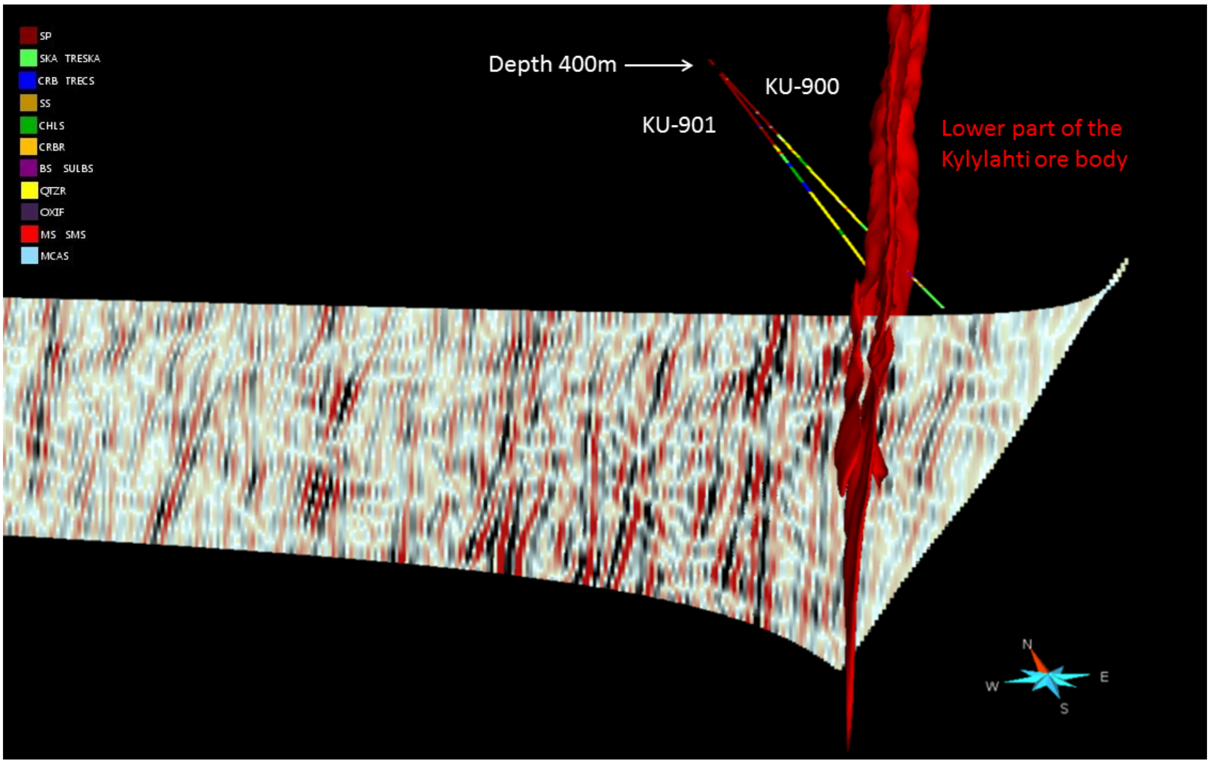

Figure 46: A VSP seismic section recorded from the drill hole KU-917. Kylylahti ore body and the drill holes KU-900 and KU-901 are also shown. Modified from (Riedel et al. 2017b).

Interpretation of VSP data in complex geological environments is challenging. The approach chosen by COGITO-MIN workgroup has been to generate synthetic versions of the acquired datasets by three-dimensional, full-waveform seismic modelling using a realistic geological input model. Comparing the synthetic and acquired data, known ore deposits and lithological contacts causing reflections can be detected, as well as new areas of interest (Riedel et al. 2017a). Density model for synthetic modelling was interpolated from the Kylylahti geophysical database (Boliden 2016b). Velocity model was based on the results of this work. A cross section of simplified geological model, used as the starting point for modelling, was shown in Figure 5. The rock units included into the model were the ore (massive and semi-massive), Outokumpu ultramafics (serpentinites and soap stones), alteration fringes (carbonate-skarn-quartz rocks) and Kalevian sediments (black schists and mica schists). The velocities for these rock units were taken after the most representative rock type of each unit, massive ore for the ore, serpentinite for ultramafics, tremolite skarn for alteration fringes, and sulphide-rich black schist for Kalevian sediments (Riedel et al. 2017b). Later, Kalevian rocks were divided to mica schist and sulphide-rich black schist and ultramafics to serpentinites and soap stones. These changes were justified both by laboratory results and results of modelling (M.Riedel 2018, pers.comm.).

Figure 46 shows several clear reflections. When starting from the recording borehole on the right, the first reflection is interpreted to be caused by the contrast between the Kalevian sedimentary rocks (mica schists and black schists) and the alteration zone of the ultra-mafic massif (carbonate-skarn-quartz rocks). The next, strong reflection is caused by the ore body. In Figure 46 this reflection is partially behind the illustrated ore. The reflection after the ore body coincides well with the contact between the alteration zone (carbonate-skarn-quartz rocks) and the ultramafics (serpentinites and soap stones) (Riedel et al. 2017b). According to petrophysical results, the impedance contrast between serpentinites and any of the carbonate-skarn-quartz rocks is not necessarily high enough to produce a detectable reflection. On the other hand, a sequence of talc carbonate rocks in between could be the reason. Also a layer of tremolitic calc-silicate rock (TRECS) could cause a reflection. And finally, these alterated margins are strongly schistose and folded. As that can cause strong anisotropy to the seismic wave velocities, the cause of the reflection could be partially, or even entirely, related to the rock structure rather than the rock type.

From these examples it is obvious that seismic reflection data has great potential for exploration work in Kylylahti and in hardrock environments in general, especially when combined with other geophysical and geological data from the survey area. Synthetic models can be a great aid when interpreting data. As models always are simplified examples of the real structures and lithologies, it is also important to know, which rock types can be clustered together for modelling purposes. For realistic modelling choosing right densities and velocities for rock units is crucial.

6.6 Suggestions for interesting topics for further studies

As the importance of seismic methods in mineral exploration is growing, the petrophysical properties, density and seismic wave velocities, are of great importance. Densities are usually well known and straightforward to measure, whereas there is much less velocity data. Therefore, more seismic velocity data are needed in order to get wider references for planning seismic surveys and interpreting data in diverse geological environments.

It is well known, that seismic wave velocities increase with increasing pressure and that the growth is fastest in moderate pressures. Such pressures coincide with the areas of interest for seismic exploration, the crust in shallow depths. Although velocities increase for all rock types, the rate of growth can be quite different. Measuring the velocities in atmospheric pressure is much easier, cheaper and faster than under pressure, but the most accurate laboratory measurements would be the ones done under in-situ pressures, From already existing data, and preferably also from a well designed new set of pressurized measurements, it should be possible to make pressure corrections. With these corrections, the velocities measured in atmospheric pressure could be adjusted to be more reliable estimates of the in-situ velocities.

Metamorphic grade is another factor affecting greatly the seismic velocities of rocks, as well as other petrophysical properties. A detailed study of the effects of metamorphism to elastic properties of rocks would give new insights to that. It would also help in applying seismic velocities measured on one site to larger areas.

7 Conclusions

The aim of this work was to deepen the petrophysical knowledge of rocks in the Kylylahti area, and especially provide parameters for seismic modelling. For this purpose a set of 239 rock samples, representing 14 rock types from Kylylahti and Outokumpu-type ores from five different sites, was measured. As the petrophysical properties of rocks are highly variable even within small areas and same rock types, the sample set cannot be considered large. It was still large enough to characterize the petrophysical properties of Kylylahti rocks, partly known from previous studies or deduced from geological knowledge, especially the metamorphic zoning of the Outokumpu district. To the seismic wave velocity data from the Outokumpu district, this work gives significant addition. Previously there was data only from the Outokumpu Deep Drill Hole. The velocities measured in this study have already been successfully used for seismic forward modelling.

The density data divides the Kylylahti rocks in three categories. Massive and semi-massive sulphide mineralizations are in their own group with an average density of 3750 kg/m^3. Outokumpu assemblage rocks have densities close to 3000 kg/m^3 and Kalevian rocks, mica schist and black schist, a bit under 2800 kg/m^3. Sulphide disseminations are common in both Outokumpu assemblage carbonate-skarn-quartz rocks and black schists elevating the densities when abundant.

P-wave velocities have a wide range of values, also within the rock types. The average velocities for almost all Outokumpu assemblage rock types, including the ore, are a bit over 6 km/s. Soap stones, mica schists and black schists have lower P-wave velocities, around 5.5 km/s.

Recommended parameters (Table 11), densities and P-wave velocities, for seismic modelling in Kylylahti were discussed. The results of this work indicate, that the sulphide mineralizations produce a detectable reflection against any background due to their high density. Also the Outokumpu assemblage rocks, forming the ultra-mafic massifs, have a clear contrast against the surrounding Kalevian rocks, mica schists and black schists, even when the black schist are rich in sulphide disseminations. Soap stones are an exception. The contact between Kalevian rocks and soap stones is hardly reflective at all, whereas soap stones in contact with other Outokumpu assemblage rocks form a reflecting contact. The contacts between the alteration fringes of ultramafic massifs, carbonate-skarn-quartz rocks, and serpentinites is not expected to be reflective on average. On the other hand, densities and P-wave velocities of carbonate-skarn-quartz rocks are highly variable. These rocks are also folded and schistose causing strong anisotropy to seismic wave velocities. It is thus possible, that these rocks form reflectors within the alterated zones or at their borders, even when the average velocities or densities do not show it. Finally, it has to be kept in mind, that whether these reflective contacts are detectable from the seismic data, depends also on geometry of the structures and survey.

Porosity of the samples was very low overall and is not a major factor on any of the physical properties of Kylylahti rocks. Porosity, or microcracking, due to core retrieval is expected to affect the results of drill core samples, especially the P-wave velocities. That effect was not clearly seen in the results. That might be due to the relatively shallow original depth of the samples. Also the long water saturation time used in the measurements procedure is expected to partially compensate for the effects of microcracks.

Most of the Kylylahti rocks belong to paramagnetic group, having susceptibilities under 2000 µSI. Serpentinites and tremolitic calc-silicate rocks have higher susceptibilities and belong to strongly magnetic group. Both have low Q ratios, thus the magnetic mineral in these rocks is most probably coarse-grained magnetite. The other rock types have quite low susceptibilities on average. However, almost all of them have several samples with high susceptibility. This is due to sulphide disseminations. These samples, representing carbonate-skarn-quartz rocks, have also high Q ratios, thus the disseminated sulphides are probably mostly monoclinic pyrrhotites. The sulphide mineralizations in Kylylahti are pyrite dominated. They have quite low susceptibilities and only a few samples with high Q ratio.

Both Kylylahti sulphide mineralizations and black schists are conductive rocks. Also carbonate-skarn-quartz rocks can contain enough sulphides to make them conductive. On the other hand, most of the sulphides are disseminated, even in ore samples, thus their conductivity is not very high. The rocks containing disseminated sulphides have high IP estimates, meaning that they are chargeable. Conductivity of the black schists is due to graphite and to some extent due to disseminated sulphides.

Comparison of physical properties of the ore samples from different mine sites in the Outokumpu district reveals the differences in their mineralogy. The physical properties change with changing proportions of pyrite, pyrrhotite and magnetite. This change also follows the metamorphic zoning in the Outokumpu district, also evidenced by the properties of Outokumpu district serpentinites. The chrysotile serpentinites in the west are on average notably lighter and have higher susceptibilities than the antigorite serpentinites in the east.

According to the physical properties of Kylylahti rock types, ultramafic massifs like Kylylahti massif can be detected from the Kalevian rocks surrounding them by gravity, magnetic and seismic methods. The sulphide mineralizations can be detected by gravity and seismic methods if they are big enough, and the geometry is favorable for their detection. Electromagnetic methods can be used for locating conductive areas, although these can be either caused by sulphide mineralizations or by black schists. As always with geophysical surveying, best results are achieved combining several methods and geological knowledge.

8 Acknowledgements

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